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American Journal of Science

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THE UNIT CELLS OF CALCITE

HORACE WINCHELL

ABSTRACT. Mineralogical literature is confused regarding the axial lengths of the morphological unit cells of crystals belonging to the calcite structure-type. The smallest unit cell is a steep rhombohedron, $\{10\bar{4}1\}$ if the cleavage is $\{10\bar{1}1\}$; the axes **a** and **c** of the corresponding hexagonal cell are about 4.99 and 17.06 A., respectively. The axial lengths usually quoted for the morphological cells, both rhombohedral and hexagonal, correspond to subcells with $\mathbf{c}' = \frac{1}{2}\,\mathbf{c}$, long enough only to extend from a CO_3-group with one corner toward the front, to a CO_3-group with a corner toward the back. The axes **a** and **c** corresponding to the cleavage rhombohedron as $\{10\bar{1}1\}$ are about 19.96 A., and 17.06 A., respectively, and the corresponding rhombohedral axes are 12.85 A. long. A complete set of transformation matrices between the hexagonal and rhombohedral cells, both structural and morphological, is in tables 1 and 2.

Errors have occurred since the structure of calcite was first compared (Bragg 1914; 1915) with that of halite by observing that the pattern of the calcium and the carbon atoms (oxygen atoms being disregarded) is similar to that of the sodium and the chlorine atoms in a halite structure deformed by compression along a cube-diagonal [111] so as to make a rhombohedron with axes $\mathbf{a}'_{Rh} = 6.42$ angstroms and interaxial angles $\alpha_{Rh} = 101°55'$. Such a deformation is reasonable because of the flat shape of the CO_3 groups, which lie with their planes parallel to (0001) and their threefold symmetry axes parallel to **c**. These triangular groups have the same orientation in alternate planes; in the intervening planes they have another orientation, turned 180° (or 60°) about **c** as compared with the groups in the first-mentioned planes. Without this second orientation of the CO_3 groups there could be no center of symmetry in calcite, and either the planes or the twofold axes of symmetry would likewise be impossible.

Dana's System of Mineralogy (Palache, Berman, and Frondel, 1951, p. 141) describes the CO_3 groups in the calcite structure as follows: "These groups are arranged in parallel position in the crystal with their threefold symmetry axes oriented along [0001]. . . ." The unfortunate words "parallel position" should of course be "parallel planes."

Elsewhere, the *System* (p. 158, ref. 1) gives a transformation matrix as follows: "morphology to structure 1000/0100/0010/0004." This is correct from the morphological viewpoint, but as will become evident below, a minor change would make it correct not only morphologically but also structurally.

Bragg (1914a,b, 1915), Schiebold (1919), Olshausen (1925), and perhaps others have apparently all erred in considering the alternate possible unit cells of calcite. Wyckoff (1920) is the first I have found who clearly and explicitly states that the unit cell with $\mathbf{a}'_{Rh} = 6.42$ and $\alpha_{Rh} = 101°55'$, useful in comparing with the halite structure, is not a unit cell of calcite,

though the idea is implicit in the work of Bragg (1915, esp. fig. 15) and of all others who found the correct primitive rhombohedral cell containing $2CaCO_3$, with $\mathbf{a}_{Rh} = 6.37$ and $\alpha_{Rh} = 46°05'$. Wyckoff (1948-1951, table VIIA, 2) may give the impression that there is a cleavage-rhombohedral cell with $\alpha_{Rh} = 101°55'$ and $\mathbf{a}'_{Rh} = 6.42$, but this is corrected in his text (Par. VII, a1) with the statement that it is a subcell. Maugin (1923) states explicitly that a rhombohedral unit cell exists with $\alpha_{Rh} = 101°55'$, but that its edge $\mathbf{a}_{Rh}$ must be twice as long as the generally adopted value of $\mathbf{a}'_{Rh}$. Ewald and Hermann (1931, p. 292, 313, 316, etc.) recognize the necessity of doubling the axes $\mathbf{a}'_{Rh}$ to obtain a unit cell of the calcite structure; their description is clear, but evidently not widely enough known. The size and shape of the structural unit cell with $\alpha_{rh} = 46°05'$ has been given by several authors, as Andrews (1950) who also summarizes previous work, without discussing its relation to, or the size of, the morphological unit cell.

The cleavage-like subcell, with $\mathbf{a}'_{Rh} = 6.42$, $\alpha_{Rh} = 101°55'$, has only one-eighth the volume and one-half the edge-length of a true lattice cell of calcite, for if the origin of coordinates is placed at a carbon atom, then each rhombohedral "axis" $\mathbf{a}'_{Rh}$ extends from the origin to a carbon atom at the center of a CO_3-group of the opposite orientation; $\mathbf{a}_{Rh}$ must be doubled to extend to a group with the same orientation as that at the origin. The doubled length, $\mathbf{a}_{Rh} = 2\mathbf{a}'_{Rh}$, is the axial length for this cell. The statement that the carbon and calcium atoms form a halite-like structure with the smaller subcell is still perfectly true, but this is not a unit cell of the whole calcite structure.

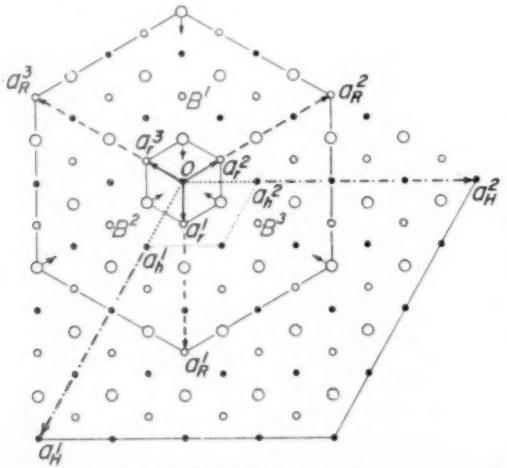

Fig. 1. Part of a rhombohedral lattice. Solid circles represent points in the lowest, or origin plane; small and large open circles represent points in the first and second lattice planes above the origin; the third plane has points directly over the solid circles, but is omitted for clarity. Axes a_r define the smallest, or structural rhombohedral unit cell; a_h (and c, not shown, normal to the page at O) define the corresponding hexagonal unit cell. Similarly, a_R and a_H (and c, not shown) define the rhombohedral and the hexagonal unit cells corresponding to the cleavage rhombohedron, as $\{10\bar{1}1\}$, of crystals with the calcite structure.

Figure 1 is a diagram of a rhombohedral lattice. By their sizes the circles represent lattice points at three different levels including, and above, the plane containing the origin O. The fourth level would have points coinciding in this figure with those of the first. The threefold symmetry axis is normal to the plane of the diagram. A primitive (structural) rhombohedral cell with axes $\mathbf{a}_{rh}$ is outlined in solid lines about O, except that its uppermost third is represented only by three arrows pointing toward the lattice point above O in the fourth level. A cell of calcite with this shape and size has parallel CO_3 groups at every corner, and an oppositely oriented CO_3 group at the center (above O, at a level of $\mathbf{c}/2$, not shown); the calcium centers are also above O, at levels (not shown) $\mathbf{c}/4$ and $3\mathbf{c}/4$. The nearest oxygen neighbors to each calcium belong to CO_3 groups associated with adjacent unit cells, so that it is hard to visualize the relations from just a single structural cell.

The hexagonal unit cell corresponding to the primitive rhombohedral cell is obtained by the transformation (1); it is said to be rhombohedrally centered; its axes $\mathbf{a}_h$ are shown by dotted lines in the figure; its axis $\mathbf{c}_h$ is normal to the page at O. The vector equations defining this cell (Henry and Lonsdale, 1952, p. 15) are as follows:

$$\mathbf{a}_1(\mathrm{h}) = \mathbf{a}_1(\mathrm{rh}) - \mathbf{a}_2(\mathrm{rh})$$
$$\mathbf{a}_2(\mathrm{h}) = \mathbf{a}_2(\mathrm{rh}) - \mathbf{a}_3(\mathrm{rh})$$
$$\mathbf{c}(\mathrm{h}) = \mathbf{a}_1(\mathrm{rh}) + \mathbf{a}_2(\mathrm{rh}) + \mathbf{a}_3(\mathrm{rh})$$

Transformation matrix, rh(struc.) to hex (struc.):

$$\left\| \begin{matrix} 1 & \bar{1} & 0 \\ 0 & 1 & \bar{1} \\ 1 & 1 & 1 \end{matrix} \right\| \quad \text{Modulus 3.} \qquad (1)$$

The lengths of the resulting axes are $\mathbf{a}_{hex} = 4.9898$, $\mathbf{c}_{hex} = 17.060$ A. This matrix and others are collected in table 1. For a discussion of many uses of these matrices, see Henry and Lonsdale (1952, p. 15).

Considering figure 1 and the requirement that each rhombohedral axis should extend from the origin to a point in the adjacent level, we find that the solid-line axes $\mathbf{a}_{rh}$ are the shortest possible. The axes OB (not drawn in figure 1) define a face-centered rhombohedral cell, and indeed they were derived by Wyckoff (1920), Olshausen (1925), and others, in the process of finding the 6.37-angstrom axes $\mathbf{a}_{rh}$, at 46°05′, that define the structural cell. The axes OB are derived from the axes $\mathbf{a}_{rh}$ by matrix (2).

$$\left\| \begin{matrix} \bar{1} & 1 & 1 \\ 1 & \bar{1} & 1 \\ 1 & 1 & \bar{1} \end{matrix} \right\| \quad \text{Modulus 4.} \qquad (2)$$

Olshausen (1925) gave the length of OB as 8.08 A., and the interaxial angle as 76°03′, essentially the same as may be calculated from the best modern data. Curiously, Olshausen seems to have overlooked the fact that simply

TABLE 1

Unit Cells and Axial Transformations in Calcite
Using 3-Axis Rhombohedral and Hexagonal Coordinates

	From rh (struc.)	From hex (struc.)	From Rh (morph.)	From Hex (morph.)
To rh (struc.)	$a_{rh} = 6.3748$ $\alpha_{rh} = 46°05'$ cell contains 2 $CaCO_3$	$\begin{Vmatrix} \frac{2}{3} & \frac{1}{3} & \frac{1}{3} \\ -\frac{1}{3} & \frac{1}{3} & \frac{1}{3} \\ -\frac{1}{3} & -\frac{2}{3} & \frac{1}{3} \end{Vmatrix}$ mod. $\frac{1}{3}$	$\begin{Vmatrix} \frac{1}{2} & \frac{1}{4} & \frac{1}{4} \\ \frac{1}{4} & \frac{1}{2} & \frac{1}{4} \\ \frac{1}{4} & \frac{1}{4} & \frac{1}{2} \end{Vmatrix}$ mod. $\frac{1}{16}$	$\begin{Vmatrix} \frac{1}{6} & \frac{1}{12} & \frac{1}{3} \\ -\frac{1}{12} & \frac{1}{12} & \frac{1}{3} \\ -\frac{1}{12} & -\frac{1}{6} & \frac{1}{3} \end{Vmatrix}$ mod. $\frac{1}{48}$
To hex (struc.)	$\begin{Vmatrix} 1 & \bar{1} & 0 \\ 0 & 1 & \bar{1} \\ 1 & 1 & 1 \end{Vmatrix}$ mod. 3	$a_{hex} = 4.9898$ $c_{hex} = 17.060$ cell contains 6 $CaCO_3$	$\begin{Vmatrix} \frac{1}{4} & -\frac{1}{4} & 0 \\ 0 & \frac{1}{4} & -\frac{1}{4} \\ 1 & 1 & 1 \end{Vmatrix}$ mod. $\frac{3}{16}$	$\begin{Vmatrix} \frac{1}{4} & 0 & 0 \\ 0 & \frac{1}{4} & 0 \\ 0 & 0 & 1 \end{Vmatrix}$ mod. $\frac{1}{16}$
To Rh (morph.)	$\begin{Vmatrix} 3 & \bar{1} & \bar{1} \\ \bar{1} & 3 & \bar{1} \\ \bar{1} & \bar{1} & 3 \end{Vmatrix}$ mod. 16	$\begin{Vmatrix} \frac{8}{3} & \frac{4}{3} & \frac{1}{3} \\ -\frac{4}{3} & \frac{4}{3} & \frac{1}{3} \\ -\frac{4}{3} & -\frac{8}{3} & \frac{1}{3} \end{Vmatrix}$ mod. $\frac{16}{3}$	$a_{Rh} = 12.850$ $\alpha_{Rh} = 101°55'$ cell contains 32 $CaCO_3$	$\begin{Vmatrix} \frac{2}{3} & \frac{1}{3} & \frac{1}{3} \\ -\frac{1}{3} & \frac{1}{3} & \frac{1}{3} \\ -\frac{1}{3} & -\frac{2}{3} & \frac{1}{3} \end{Vmatrix}$ mod. $\frac{1}{3}$
To Hex (morph.)	$\begin{Vmatrix} 4 & \bar{4} & 0 \\ 0 & 4 & \bar{4} \\ 1 & 1 & 1 \end{Vmatrix}$ mod. 48	$\begin{Vmatrix} 4 & 0 & 0 \\ 0 & 4 & 0 \\ 0 & 0 & 1 \end{Vmatrix}$ mod. 16	$\begin{Vmatrix} 1 & 1 & 0 \\ 0 & 1 & \bar{1} \\ 1 & 1 & 1 \end{Vmatrix}$ mod. 3	$a_{Hex} = 19.959_2$ $c_{Hex} = 17.060$ cell contains 96 $CaCO_3$

Notes: 1. Structural elements have lower-case subscripts, morphological elements have capitalized ones.

2. The modulus of each matrix is the ratio of volumes of new to old cells.

doubling the length of the "axes" $\mathbf{a'}_{Rh}$ would give a valid unit cell. A hexagonal cell is derivable from the rhombohedral one with axes OB using the transformation matrix (1); this has axes $\mathbf{a'}_h$ twice as long as, and directed in the opposite sense to, the axes $\mathbf{a}_h$ of the structural hexagonal cell.

Another rhombohedral cell in the same series is obtained from the axes

TABLE 2

Axial Transformations in Calcite

Using 3-Axis Rhombohedral, and 4-Axis Hexagonal Coordinates

	From rh (struc.)	From hex (struc.)	From Rh (morph.)	From Hex (morph.)
To rh (struc.)	*	$\begin{Vmatrix} \frac{1}{3} & 0 & -\frac{1}{3} & \frac{1}{3} \\ -\frac{1}{3} & \frac{1}{3} & 0 & \frac{1}{3} \\ 0 & -\frac{1}{3} & \frac{1}{3} & \frac{1}{3} \end{Vmatrix}$	*	$\begin{Vmatrix} \frac{1}{12} & 0 & -\frac{1}{12} & \frac{1}{3} \\ -\frac{1}{12} & \frac{1}{12} & 0 & \frac{1}{3} \\ 0 & -\frac{1}{12} & \frac{1}{12} & \frac{1}{3} \end{Vmatrix}$
To hex (struc.)	$\begin{Vmatrix} 1 & \bar{1} & 0 \\ 0 & 1 & \bar{1} \\ \bar{1} & 0 & 1 \\ 1 & 1 & 1 \end{Vmatrix}$	*	$\begin{Vmatrix} \frac{1}{4} & -\frac{1}{4} & 0 \\ 0 & \frac{1}{4} & -\frac{1}{4} \\ -\frac{1}{4} & 0 & \frac{1}{4} \\ 1 & 1 & 1 \end{Vmatrix}$	$\begin{Vmatrix} \frac{1}{4} & 0 & 0 & 0 \\ 0 & \frac{1}{4} & 0 & 0 \\ 0 & 0 & \frac{1}{4} & 0 \\ 0 & 0 & 0 & 1 \end{Vmatrix}$
To Rh (morph.)	*	$\begin{Vmatrix} \frac{4}{3} & 0 & -\frac{4}{3} & \frac{1}{3} \\ -\frac{4}{3} & \frac{4}{3} & 0 & \frac{1}{3} \\ 0 & -\frac{4}{3} & \frac{4}{3} & \frac{1}{3} \end{Vmatrix}$	*	$\begin{Vmatrix} \frac{1}{3} & 0 & -\frac{1}{3} & \frac{1}{3} \\ -\frac{1}{3} & \frac{1}{3} & 0 & \frac{1}{3} \\ 0 & -\frac{1}{3} & \frac{1}{3} & \frac{1}{3} \end{Vmatrix}$
To Hex (morph.)	$\begin{Vmatrix} 4 & \bar{4} & 0 \\ 0 & 4 & \bar{4} \\ \bar{4} & 0 & 4 \\ 1 & 1 & 1 \end{Vmatrix}$	$\begin{Vmatrix} 4 & 0 & 0 & 0 \\ 0 & 4 & 0 & 0 \\ 0 & 0 & 4 & 0 \\ 0 & 0 & 0 & 1 \end{Vmatrix}$	$\begin{Vmatrix} 1 & \bar{1} & 0 \\ 0 & 1 & \bar{1} \\ \bar{1} & 0 & 1 \\ 1 & 1 & 1 \end{Vmatrix}$	*

* See corresponding box in table 1.

OB by using matrix (2), but it can be obtained directly from $\mathbf{a}_{rh}$ by the transformation rh(struc.) to Rh(morph.):

$$\begin{Vmatrix} 3 & \bar{1} & \bar{1} \\ \bar{1} & 3 & \bar{1} \\ \bar{1} & \bar{1} & 3 \end{Vmatrix} \quad \text{Modulus 16.} \qquad (3)$$

This cell has axes $\mathbf{a}_{Rh} = 12.85$ A., designated in figure 1 by $\mathbf{a}_R$, drawn with dashed lines, at interaxial angles $\alpha_{Rh} = 101°55'$. Matrix (3) will be found in table 1 in the column "From rh(struc.)" and the row "To Rh(morph.)." Application of matrix (1) to this cell gives the hexagonal-morphological cell with $\mathbf{a}_{Hex} = 19.959$ A., $\mathbf{c}_{Hex} = 17.060$ A.

Table 1 contains the transformation matrices for changing either way between rhombohedral and hexagonal, and between the smallest (structural) and the morphologically preferred unit cells, except that in squares where the identity-matrix would be appropriate, the values of **a** and α or of **a** and **c** are shown.

Table 2 contains the same transformations, but with reference to four-axis Bravais coordinates for all hexagonal members.

CONCLUSION

Needless confusion regarding the axial lengths for the morphological unit cells of crystals belonging to the calcite structure-type has arisen. Axial lengths in any structure are the true repeat-distances for the whole structural pattern; the lengths of the halite-like rhombohedral subcell-edges are not axial lengths; they must be doubled when the positions of the oxygens in the CO_3-groups are considered. Likewise the lengths of the corresponding hexagonal subcell edges are not axial lengths, but must be doubled for the same reason. In all discussions involving the structure or any vector properties related to crystallographic axial lengths, authors should carefully state the reference-axes used in their work.

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INTERPRETATION OF FABRICS OF EXPERIMENTALLY DEFORMED SANDS

IRIS Y. BORG and JOHN C. MAXWELL

ABSTRACT. Quartz sands deformed at 26,000-32,000 p.s.i., 270°-320° C., for periods of 18 hours to 100 days in the presence of slightly alkaline solutions experience a reduction in porosity and permeability and are markedly coherent. Slight reduction in the size of grains, interlocking grain boundaries, and extensive fracturing are conspicuous features. $r(10\bar{1}1)$ and $z(01\bar{1}1)$ are the most important surfaces of rupture and make up 44% of all fractures; planes in the prism zone, particularly $m(10\bar{1}0)$, make up 18% of the remaining breaks. Fractures occupy cones approximately 15° from the axis of compression and are interpreted as tension fractures. Grains of the deformed sand are dominantly elongated parallel to the c axis and r and z. A strong dimensional orientation of grains was produced, and a weak preferred orientation of c axes is a concomitant feature. Petrofabric diagrams are similar to those of natural bedded sandstone.

INTRODUCTION

For the past five years Maxwell has been carrying out a study of the effects of elevated temperature and pressure on the consolidation of sands. Cementation, compaction, and porosity have been investigated under conditions simulating burial at depths of 16,000 to 35,000 feet. The study has been extended to include behavior of quartz grains under directed stress, the subject of the present report. Notable contributors to knowledge of behavior of quartz under controlled stress are Griggs (1936), Griggs and Bell (1938), and Fairbairn (1950). The purpose of these investigations has been to shed some light on the mechanisms by which quartz attains a preferred orientation in tectonites, and it was with this ultimate goal in mind that the present analyses were undertaken.

Although the material used in Maxwell's experiments was derived from several sources, the fabric analyses are restricted to deformed specimens of an unconsolidated beach sand from Asbury Park, New Jersey. The range of experimental conditions represented by the ten specimens is as follows:

Temperature: 270°-320° C.
Confining pressure: 11,500-16,000 p.s.i.
Directed pressure: 26,000-32,000 p.s.i.
Duration of test: 18 hours to 100 days
Solutions: Seawater, pH 8.0, and 5% Na_2CO_3 solution

EXPERIMENTAL TECHNIQUE

A detailed description of apparatus and experimental technique has already been published (Maxwell and Verrall, 1954), and the subject will be only briefly treated. The sand, which is saturated with seawater, is poured into a metal cylinder, one inch in diameter and three-eighths of an inch high, and tapped down. During the deformation confining pressures are maintained by forcing solutions into the specimen and directed stress by applying a piston to the cylinder. First directed pressure and then confining pressure are applied; the temperature is raised to the desired level; and finally the pressures are raised to the values at which the experiment is to be run.

The small size of the cylinders together with the friable nature of the material limited to one the number of thin sections that could be successfully cut. All measurements of *c* axes, lamellae, fractures, elongations, and planes of inclusions are taken from sections cut parallel to the axis of compression. Since axial symmetry is anticipated in any orientation produced by the stress, no difficulties are encountered by restricting measurement to a single section. Vacancies in petrofabric diagrams because of "blind spots," e.g. areas representing planes that are inaccessible, must be interpreted in terms of the required symmetry. A five-axis Leitz universal stage and a Schmidt equal-area projection were used throughout.

ASBURY PARK SAND

The Asbury Park sand is a medium-grained (average grain diameter, 0.28 mm), well rounded and sorted, pure quartz sand. Approximately 60% of the grains show some sign of fracturing; of these 42% are slightly fractured, 10% moderately fractured, and 8% highly fractured. Metamorphic features such as undulatory extinction, deformation lamellae, and deformation bands have been inherited from previous deformation. Undulatory extinction is visible in 59% of the grains; 15% are strongly undulatory, 18% moderately, 26% slightly, and 41% show very little or none. Deformation lamellae are present in 20 out of 200 grains or in 10% of the grains. As many as three separate sets were observed in a single grain, and two sets per grain are not uncommon.

A large percentage of the grains contain fluid or opaque inclusions that tend to be aligned in planes. It is difficult to decide whether their formation attended the initial crystallization of the quartz or whether they owe their existence to subsequent fracturing and annealing in the presence of solutions as postulated by Tuttle (1949). 40% of the 159 planes of inclusions measured in an 80-grain sample make angles of 0°-15° with the *c* axis. None were recorded 75°-90° to the *c* axis, i.e. subparallel to the basal plane. 60% are dispersed over the remaining interval, 15°-75° to the *c* axis with no conspicuous maxima.

Tuttle (1949, p. 354) in a study of liquid inclusions concludes that "—there is no control of the orientation of the planes by crystallographic directions in quartz suggesting that quartz behaved as an isotropic material." Yet comparison of his petrofabric diagrams of *c* axes and poles to planes of inclusions within the same specimen reveals a tendency in five out of seven examples for the poles to lie at 90° to the *c* axis maxima. The preference of the planes of inclusions to lie parallel to the *c* axis is in keeping with observations to be presented that quartz under metamorphic conditions does not behave as an isotropic material.

BEHAVIOR OF DEFORMED SAND

General remarks.—Table 1 is a summary of the specific experimental conditions attending the deformation of individual specimens.

TABLE 1
Experimental Conditions

Specimen	Temperature ° C.	Confining pressure p.s.i.	Directed pressure p.s.i.	Duration of test	Nature of solutions
I - 3	320°	16,000	32,000	18 hours	5% Na_2CO_3 pH 11.4
I - 6	270°	11,500	26,500	44 hours	Seawater pH 8.0
I - 7	300°	13,000	29,000	10 days	"
II - 1	280°	"	"	93 hours	"
III - 1	285°	"	"	100 days	"
IV - 1	320°	"	"	7 days	"
IV - 9	285°	"	"	26 hours	"
IV - 12	285°	"	"	24 hours	"
IV - 30	285°	"	"	5 days	"
IV - 35	300°	"	"	100 days	"

In all specimens the most obvious deformational feature is fracturing. A second, less conspicuous feature of the deformed material is a dimensional orientation of grains. A close examination reveals interlocking grains, radially aligned fractures at grain contacts, and in addition undulatory extinction and lamellae, which are present in the original material. Neither cementation nor growth of grains is conspicuous in thin section, but they may be inferred from the large changes in porosity and permeability (Maxwell and Verrall, 1954) and the coherency of some of the specimens deformed for periods exceeding ten days. It is surprising that recrystallization effects are not more obvious as quartz has been grown from 5% Na_2CO_3 solutions saturated with SiO_2 at a rate of 0.1 inch per day at approximately 380°C. and 15,000 p.s.i. (Buehler and Walker, 1949). Experimental growth of quartz is characterized by the presence of thermal and solubility gradients within the bomb thereby facilitating the transport of silica. The apparatus used in the deformations was designed to yield the greatest possible uniformity of temperature over the length of the specimen, hence the most favorable conditions for transport and redeposition of silica are not present. Local operation of the mechanism of solution and redeposition described by Riecke's principle probably accounts for the cementation that does occur.

Dimensional alignment of grains.—Photomicrographs show that grains tend to be elongated in a particular direction within each thin section. An analysis of the azimuthal distribution of the direction of elongation in mutually perpendicular sections showed that the longest dimension of the grains tends to lie near the plane normal to the axis of compression. Figure 1 shows the distribution of long axes as compiled from five specimens and measured in sections cut parallel to the axis of compression. Approximately 10% of the grains encountered in each section were disregarded because they appeared equidimensional or nearly so. The direction of elongation of 47% of the 534 grains measured is within 30° of the plane normal to the axis of

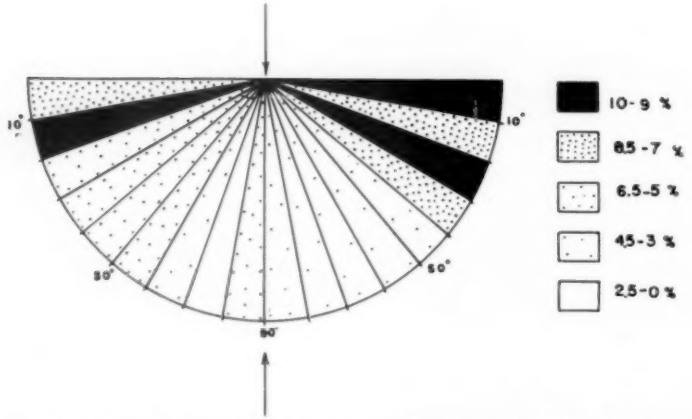

Fig. 1. Azimuthal distribution of long axes of quartz grains. 534 grains were measured in 5 thin sections. Plane of paper corresponds to the plane of the thin section. Arrows indicate position of axis of compression.

compression. The dimensional orientation is produced during packing of the cylinders by a mechanism similar to that effecting dimensional orientation in bedded sands (Wayland, 1939; Griffiths and Rosenfeld, 1950). To what extent mechanical rotation induced by the deformation contributes to the alignment has as yet not been determined.

Relation of elongation to the c axis.—The direction of elongation of a grain as measured in thin section is the trace of the direction of the greatest elongation except in those rare instances where this direction is nearly normal to the plane of the section. In thin section analysis of grain shape the angle between the trace of the direction of greatest elongation and the *c* axis of a grain is meaningless unless the direction of greatest elongation lies very near the plane of the thin section. A more informative relation is the angle between the trace of the longest dimension and the trace of the *c* axis, whose maximum value is set by the angle between the *c* axis and the direction of greatest elongation. Figure 2 is a graphical representation of the angles observed between the traces of the *c* axis and elongation as measured in thin section. The data indicate that the grains are dominantly elongated parallel to the *c* axis and the rhombehedrons, *r* and *z*, 38° from the *c* axis. The results are in keeping with the observations made on natural sandstones by Rowland (1946) and on metamorphic rocks by Ingerson and Ramisch (1942).

It was noted that deformation tends to destroy small grains and to modify the shape of the large grains. Since the small grains are pulverized, elongations recorded in deformed material are presumably inherited from those of the larger grains of the undeformed sands. That this is so is brought out particularly well by measurement of grain size in deformed material. In specimens deformed for five to seven days there is a tendency for the average grain size to increase over that recorded in the undeformed sand. For example, by using a point counter the average grain dimension in Specimen IV-1, de-

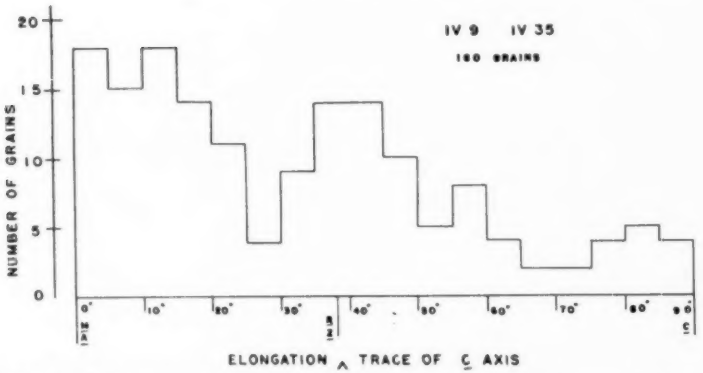

Fig. 2. Relation between observed direction of elongation and trace of the *c* axis.

formed seven days, was determined as 0.33 mm parallel to the axis of compression and 0.37 mm normal thereto. The average grain size in the unoriented and undeformed loose sand is 0.28 mm. In Specimen IV-35, deformed one hundred days, the average parallel to the axis of compression is 0.27 mm and at right angles to the axis 0.29 mm. The results graphically represented in figures 1 and 2 describe the elongation and grain shape of the slightly modified large grains in the undeformed sand.

Orientation of c axes.—Separate plots were made of *c* axes in fractured and unfractured grains, but no consistent difference in orientation could be detected. One feature is common to both, a tendency for the *c* axes to lie at high angles to the axis of compression. Figure 3A is a compilation of *c* axes measured in six thin sections. At first glance the weak concentration normal to the axis of compression may appear of no significance; it is manifest however in all individual sections. Figure 3B is a composite of maxima from the individual slides. The weak girdle stems from the dimensional elongation of grains recorded earlier in this paper.

Nature of the fractures.—In all specimens regardless of the length of time of the test, the percent of grains still intact and showing a high to moderate degree of fracturing ranges from 50 to 60%, as contrasted to 18% in the undeformed material. Fractures are conjugate in most cases, i.e. they are breaks parallel to two or more planes that have joined to produce a single discontinuity. The same conclusion was reached by Anderson (1945). Measurement of the planes of discontinuity is difficult; nonetheless if attention is focused on planar portions of the rupture, identification is usually possible. A detailed study of fractures was made in one hundred grains of Specimen IV-1. The *c* axis and poles to all planar fractures occurring in each grain were measured and plotted on a projection. Angular and zonal relationships were utilized to identify the fractures. The results are tabulated below.

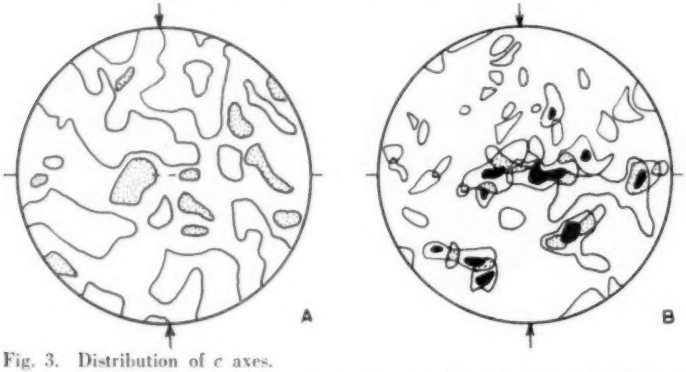

Fig. 3. Distribution of *c* axes.

A. 942 *c* axes from 5 deformed specimens (6 thin sections). All data have been rotated to bring axes of compression (arrows) into coincidence. Contours: 0-1-1.5%, maximum 2%.

B. Superimposition of maxima from each of five contoured *c* axes petrofabric diagrams.
Total of 900 *c* axes measured in four deformed specimens. Solid areas represent 3% maxima, blank areas, 2% maxima, and dotted areas, overlap of 2% maxima.

TABLE 2

Identification of Fractures

	$r(10\bar{1}1)$ and $z(01\bar{1}1)$	$m(10\bar{1}0)$	$a(11\bar{2}0)$	$s(11\bar{2}1)$	$\xi(11\bar{2}2)$	$c(0001)$	$d(10\bar{1}2)$
No. of times encountered	103	26	10	10	6	2	7
Number of unidentified fractures			80				
Total	249 in 100 grains						

Among the unidentified fractures some are undoubtedly parallel to the prisms, *m* and *a*; in the absence of other identifiable planes, however, they could not be distinguished. Others are probably *r* or *z*, but because they were recorded as occurring more than 5° away from the accepted position, they were not labeled. The histogram of figure 4 shows the position of all fractures relative to the *c* axis in the one hundred grains whether identified or not.

On the basis of the angle between the direction of greatest elongation and *c* axis, it has been assumed that the rhombehedrons, *r* and *z* are the most important planes of rupture in quartz. The results recorded here confirm the deduction. Next in importance are the prisms. The basal plane plays little part in the fracturing. It might be anticipated that fracturing would be governed to a large extent by the prisms for two reasons: (1) Undulatory bands, which are common in the undeformed sand, are commonly bounded by microfractures parallel to the prism faces. (2) Planes of inclusions, also common in the sands, are planes of weakness; and they show a preference for the prism zone. Fractures crossing the zones of undulatory extinction and planes of inclusions are not uncommon.

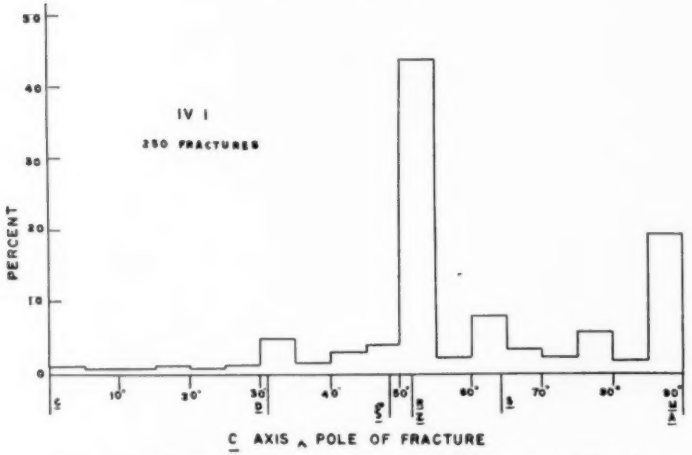

Fig. 4. Crystallographic position of fractures. 250 fractures in 100 grains.

Spatial distribution of fractures.—Fractures frequently radiate from points of contact of neighboring grains (fig. 5) and indicate that stress is transmitted at grain contacts. Sets of parallel trending fractures are rare, and as many as six breaks trending in different directions can be measured within a single grain. A plot of the poles of fractures in each thin section consistently produced double maxima about a normal to the axis of compression. A compilation of fractures occurring in 450 grains is shown in figure 6. The compilation was made by superimposing the axis of compression and the plane of the thin section in each of five specimens. Since there is a blind spot in each of the sections, the diagram of figure 6 reflects a double girdle, in keeping with the axial symmetry of the deformation. The fractures are interpreted as tension breaks. They depart from the ideal position of a tension fracture by an average of 15° and in doing so are reminiscent of *ac* joints which rarely occur at exactly 90° to the fold axis.

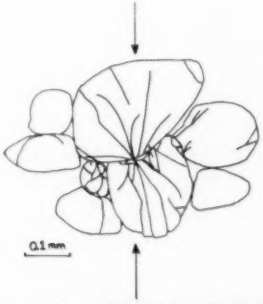

Fig. 5. Drawing of grain contact illustrating development of radial fractures and interlocking grains.

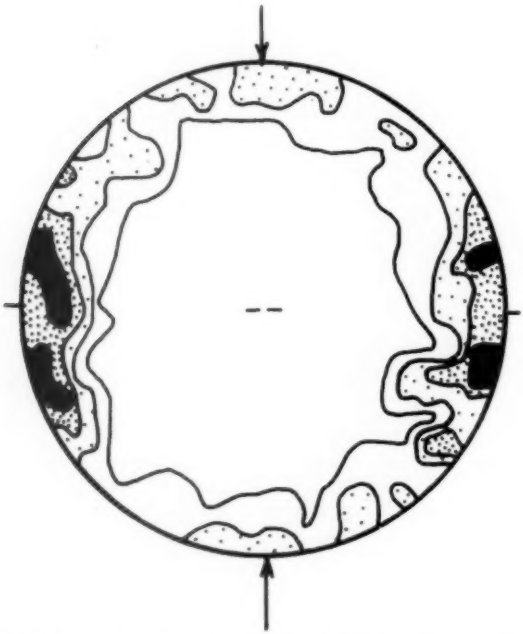

Fig. 6. Areal distribution of fractures. Poles of 919 fractures measured in 450 grains in five deformed specimens. All measurements have been rotated to bring axes of compression (arrows) into coincidence. Fractures in center are inaccessible. Contours: 0-1-1.5-2-2.5%, maximum 3%.

Undulatory extinction.—As far as could be determined, little or no undulose extinction was produced in the quartz by deformation. Actual counts in deformed material are not reliable because of extensive fracturing and slight rotation of fragments, which in many cases simulates undulatory extinction.

Deformation lamellae.—The percentage of grains containing lamellae sharply decreases in the deformed sands. In contrast to the 10% of grains containing lamellae in the undeformed sands, only 3% are present in the deformed sands. The number recorded in any one section could not be related to the length of time of the test, but if there were such a relationship, it probably could not be recognized since the number of grains containing lamellar structure within a deformed specimen is small. The possibility that grains containing lamellae survived because stress could be relieved by gliding on pre-existing planes was investigated, but it could not be substantiated. Table 3 is a tabulation of the angle the axis of compression makes with the pole to surviving lamellae in 63 examples.

TABLE 3
Position of Lamellae within Deformed Specimens

Number of lamellae in specimens deformed less than 5 days	2	2	3	4	5	1	2	3	3
Angle between axis and pole of lamellae	0°-10°	11°-20°	21°-30°	31°-40°	41°-50°	51°-60°	61°-70°	71°-80°	81°-90°
Number of lamellae in specimens deformed more than 5 days	1	2	4	4	5	6	6	7	3
Total	3	4	7	8	10	7	8	10	6

There is a tendency for the poles to lie 41° to 80° from the axis of compression in the material deformed for periods exceeding five days, but attaching any particular significance to it is unwarranted. Grains whose lamellae fall within the 41°-80° range have no particular *c* axis orientation. A characteristic of all grains containing lamellae is absence of extensive fracturing. This feature hampered attempts to describe the position of the lamellae in individual grains, for unless fractures paralleling the rhombehedrons are present, a unique orientation of the lamellae in the grain can not be established. In the seventy-eight lamellae studied, the angle between the *c* axis and the pole to the lamellae varied from 6° to 82° with a maximum at 15° to 20°. Thirty-eight sets of lamellae occurred in grains whose edges or fractures could be identified. In the latter group, the lamellae were found to occur in variable positions with a slight tendency, in fourteen out of thirty-eight cases, to fall in the [11$\bar{2}$0] zone.

Summary and conclusions.—(1) Reduction in average size of grains is demonstrable in specimens deformed for more than seven days. Analyses of absolute grain elongation indicate that the smaller grains are the first to break down; they are followed by the larger ones.

(2) It is questionable whether either undulatory extinction or deformation lamellae were produced under the experimental conditions attending the deformation. The percentage of grains containing lamellae decreased, as judged by the percentages recorded in the undeformed sands.

(3) A dimensional orientation of grains can be recognized readily. 32% of the grains are aligned with their long axes within 20° of the plane normal to the axis of compression, and 47% of the grains are aligned within 30° of that plane.

(4) The dimensional orientation of grains is a "lattice orientation" as well, since the angle between the direction of greatest elongation and the *c* axis was statistically shown to be most frequently near 0° and 38°, indicating that the grains are elongated parallel to the *c* axis and the rhombehedrons,

$r(10\bar{1}1)$ and $z(01\bar{1}1)$. The combined dimensional orientation and specific grain shape give rise to a weak preferred orientation of *c* axes. A weak girdle perpendicular to the axis of compression is present in all *c* axes diagrams. The extent to which the deformation has contributed to the *c* axis orientation has not as yet been determined. The slight orientation observed is attributed largely to packing of the cylinders prior to deformation.

(5) Approximately 55% of all grains of a deformed specimen are highly to moderately fractured as contrasted to 18% showing the same degree of fracturing in the undeformed sand. Fractures rarely occur in parallel sets and with rare exception are made up of more than one crystallographic plane. If attention is focused on portions of the break most closely approximating planar surfaces, they may be identified; $r(10\bar{1}1)$ and $z(01\bar{1}1)$ are the most important surfaces of rupture and make up 44% of all recognizable planes. $m(10\bar{1}0)$ is the next most important surface and, together with other planes parallel to the *c* axis, makes up 19% of the observed fractures. Breaks within 30° of $c(0001)$ are very rare, and this investigation would indicate that they should not form any part of a theory of quartz orientation based on fracturing.

(6) Fractures occupy a cone approximately 15° from the axis of compression. This is seen in the patterns that the poles to the fractures make about the axis. They are interpreted as tension cracks which depart slightly from the ideal position.

(7) No consistent difference was observed between the *c* axis patterns of fractured and unfractured grains, indicating that the differences in strength between the directions paralleling the *c* axis and the normal thereto have no measurable reflection in the deformations described.

The conditions under which the deformations have been carried out closely resemble natural conditions to be expected at depths of three to five miles in a geosynclinal environment. It has been shown that the fabric of the sands, meaning the sum of the characteristics of its participating grains, is in large part controlled under these conditions by features initially present. In the laboratory experiments the dimensional orientation, induced by packing of cylinders, and the grain shape of the surviving grains have only been slightly modified by the simulated "overburden." The length of time of deformation materially affects the degree of porosity and permeability, but it is not easily related to the degree of orientation of either *c* axes or fractures. It can be deduced that under natural conditions the original bedding would tend to dominate the fabric of deeply buried sands.

The hypotheses of quartz orientation in tectonites are based on mechanical movement and rotation of elongate grains that tend to be bounded by specific faces (Griggs and Bell, 1938; Fairbairn, 1939). From the summary above it can be seen that the *c* axis orientations recorded here have been explained in such terms. In contrast to the assumptions made by Griggs, Bell, and Fairbairn, however, it has not been necessary to call upon grain shapes markedly different from those encountered in a beach sand. Certainly more efficient processes producing dimensional orientation can be imagined than simply tapping quartz grains down in a cylinder. It appears likely that these

same sands could give rise to varied orientations of *c* axes in other kinematic environments through direct componental movements involving rotation as well as stronger orientations than those recorded here.

ACKNOWLEDGMENTS

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WISCONSIN DRIFTS IN THE ELMIRA REGION, NEW YORK, AND THEIR POSSIBLE EQUIVALENTS IN NEW ENGLAND*

CHARLES S. DENNY

ABSTRACT. In the Elmira region, New York, Olean and Valley Heads drifts are present. Olean drift is probably older than the Wisconsin drift of southern New England, and the Valley Heads moraine in the Elmira region is probably equivalent to Rich's "late Wisconsin terminal moraine(?) stage" in the Catskill Mountains and to the prominent moraines on Long Island and Cape Cod. The correlation is based on a consideration of the topographic expression of the drifts, their content of water-laid materials, and the evidences of periglacial frost action. In the Elmira region, Binghamton drift is absent, apparently because it was completely buried by the Valley Heads ice sheet.

INTRODUCTION

In northeastern United States it has long been accepted that all the drift near the limit of the Wisconsin stage was deposited by the same ice sheet. Flint (1953) thought it likely that the terminal moraine of the Wisconsin stage (fig. 1-A) in eastern Pennsylvania and New Jersey, the two prominent moraines on Long Island, and the moraines of Cape Cod and the islands to the south mark the southern limit of an "Iowan-Tazewell complex(?)." MacClintock (1954) held that these moraines probably belong to the Tazewell substage (fig. 1-B) except for a short segment of the terminal moraine of the Wisconsin stage in northwestern New Jersey, assigned to the Cary substage. In addition, these geologists tentatively recognized in parts of New York State and in southern New England an inner drift border of the Cary substage.

Field work in the Elmira region suggests a third interpretation (fig. 1-C)—that the Olean drift (MacClintock and Apfel, 1944) is older than the Wisconsin drift in southern New England and that the Valley Heads moraine of Fairchild (1932) may be the equivalent both of Rich's (1935) "late Wisconsin terminal(?) moraine stage" in the Catskill Mountains and of the prominent moraines on Long Island and Cape Cod.

SUBDIVISIONS OF THE WISCONSIN STAGE

The Wisconsin drift of the northern part of the Elmira region was described by Tarr in the Watkins Glen-Catatonk folio (Williams, Tarr, and Kindle, 1909) as the deposits of one ice advance. He recognized a prominent recessional moraine that extends across the area just south of the Finger Lakes. In 1932 Fairchild applied the name "Valley Heads" to this moraine. The Wisconsin drift in the Salamanca Reentrant near Olean, N. Y., about 100 miles west of Elmira, was subdivided in 1944 by MacClintock and Apfel into three substages—Olean, Binghamton, and Valley Heads. The oldest drift, the Olean, forms the limit of the Wisconsin stage in eastern Pennsylvania (Leverett, 1934); the youngest drift they correlate with the Valley Heads moraine of Fairchild (1932). Their intermediate drift, the Binghamton, was established on the basis of stratigraphic, lithologic, and topographic evidence. MacClintock and Apfel traced the limit of Binghamton drift eastward from the vicinity of

* Publication authorized by the Director, U. S. Geological Survey.

Olean across the Elmira region to the western slope of the Catskill Mountains (fig. 1-B).

DRIFT SHEETS IN THE ELMIRA REGION, N. Y.

The writer believes that there are but two Wisconsin drifts in the Elmira region (fig. 1-C), the Olean of MacClintock and Apfel (1944), and the Valley Heads of Fairchild (1932). In this region the drift called Binghamton by MacClintock and Apfel (1944; MacClintock, 1954) is believed by the writer to be part of the Olean.

Olean drift.—In the Elmira region (fig. 1-C) the Olean drift covers the area south of the Valley Heads moraine of Fairchild (1932). Most of the drift contains very few erratics compared with the adjacent Valley Heads drift. Olean drift is thin, and the landscape little modified by glaciation. Terraces or benches on many slopes are controlled by bedrock. Many lower valley walls have smooth gentle slopes, and streams near their headwaters flow on very narrow flood plains (Denny, 1951, fig. 7). Many large valleys have broad alluvial fans and numerous kame terraces. In smaller valleys, deposits of water-laid drift are scarce. The tops of many kame terraces gradually merge with the adjacent valley wall along a smooth curve that is concave upward. Some glaciofluvial deposits do not form terraces but are mantled by surficial debris from the adjacent valley wall. This mantle, largely rubble of Wisconsin age, is a periglacial deposit formed by mechanical weathering and carried downslope by mass movements. The top of this mantle slopes toward the valley, and the mantle buries the initial flat top of the underlying kame terrace.

Valley Heads drift.—The drift of this substage lies north of the southern limit of Fairchild's Valley Heads moraine and contains abundant erratics of igneous and metamorphic rock (Holmes, 1952). Glaciofluvial deposits are abundant as compared with those in Olean drift. A surficial mantle of rubble, so characteristic of the region south of the moraine, is thin or absent on Valley Heads drift. The topography is more obviously affected by glaciation than to the south. Many hill slopes are not smooth but are broken by knolls, ridges, and undrained depressions. In detail many valley walls have complex slopes inclining in various directions, not smooth single slopes as in the area south of the moraine. In many valleys, flat-topped kame terraces abut sharply against the valley side and are indented by kettles. The centers of these valleys are flood plains or large kettle lakes.

The Valley Heads moraine, named by Fairchild (1932), was described in detail by Tarr (Williams, Tarr, and Kindle, 1909) in the Watkins Glen-Catatonk Folio (see also Holmes, 1952). In the north-trending through valleys south of the Finger Lakes the moraine consists of large, steep-sided hills of drift, largely water-laid, that extend north and south for several miles. The surface is dotted by small ponds and swamps. To the north in many valleys the moraine descends steeply toward the Finger Lakes. The deposits are hundreds of feet thick, but what part of the moraine was deposited by Valley Heads ice, and what part was deposited during earlier glaciations is unknown. The southern margin of the moraine is generally not prominent topographically. Its surface descends a few tens of feet to merge with a valley train that

FIG. 1A FLINT 1953

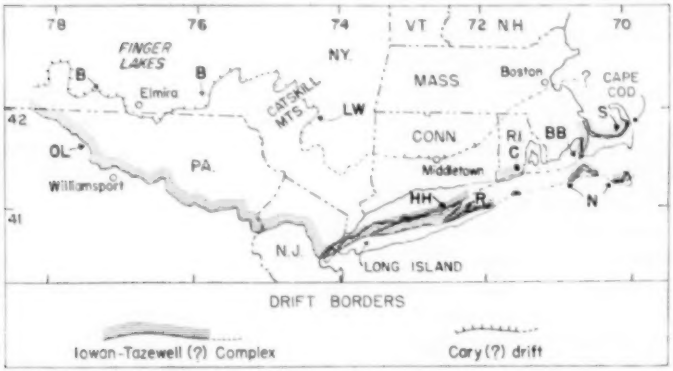


FIG. 1B MacCLINTOCK 1954

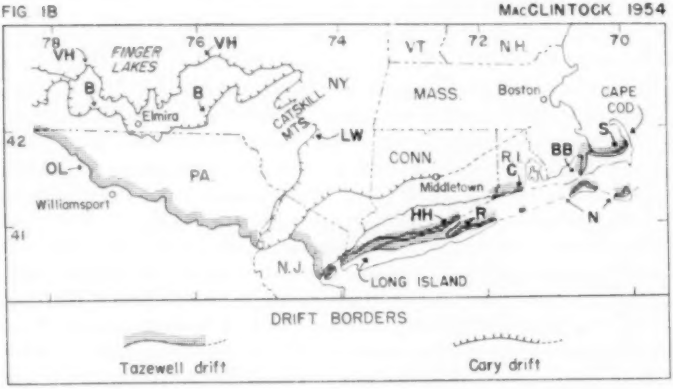
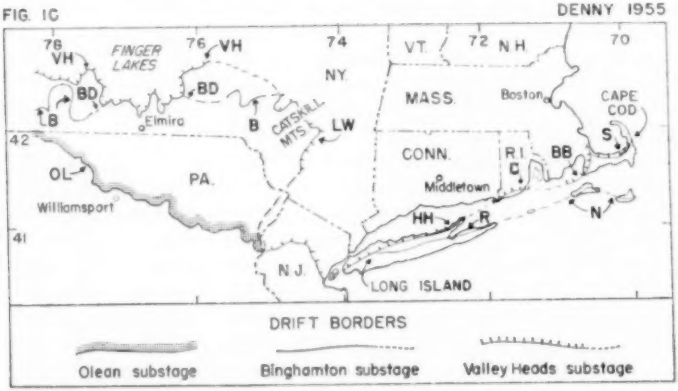


FIG. 1C DENNY 1955



SCALE
50 0 50 Miles

may extend far to the south. The moraine is the divide along much of the northern rim of the Susquehanna drainage basin. Steep-sided drainage channels, largely abandoned, are found on the sides of some valleys in the morainal belt. On the uplands between the through valleys the moraine is a distinct belt of constructional topography consisting of knolls, ridges, and closed depressions.[1]

Absence of Binghamton drift.—In the Elmira region the major east-trending through valley of the Chemung and Susquehanna Rivers (not shown on figure 1 but located just north of the border of Binghamton drift, fig. 1-B) contains drift with abundant erratics, chiefly as kame terraces. These kame terraces were used by MacClintock and Apfel (1944) to mark the southern limit of their Binghamton drift. The uplands and most of the smaller valleys to the north, however, are mantled by drift with the characteristic lithology and topographic expression of Olean drift. This anomaly suggests either that there is no Binghamton drift in the Elmira region or that here the Binghamton drift does not in most places have its characteristic topographic expression and content of erratics. No exposure of Binghamton drift resting upon Olean drift has been found here. MacClintock and Apfel (1944, p. 1157) noted Binghamton outwash gravel on Olean till 3 miles south of Corning (not shown on fig. 1-C), but the writer was unable to observe superposition of drifts at that place. However, in part of the area inside the Binghamton drift border (fig. 1-B) southwest and southeast of the Finger Lakes, the drift does have the high erratic content and topographic expression characteristic of the type area (MacClintock and Apfel, 1944). Possibly the border of Binghamton drift enters the area shown in figure 1-C from the west, passes under Fairchild's Valley Heads moraine southwest of the Finger Lakes (near Bath, N. Y., about 35 miles northwest of Elmira) and emerges again southeast of the Finger Lakes (east of Ithaca, N. Y., about 30 miles north-northeast of Elmira).

[1] The moraine is well shown on topographic quadrangle maps (scale 1:24,000; contour interval 10 feet) issued by the U. S. Geological Survey; for example, the Alpine, Mecklenberg, Ithaca West, and West Danby sheets that cover an area southwest of Ithaca, N. Y.

Fig. 1. Maps showing correlations of drift borders of Wisconsin stage in parts of northeastern United States. Figure 1-A is that suggested by Flint (1953, pl. 2); figure 1-B is that proposed by MacClintock (1954, fig. 2); figure 1-C is that of Denny. Pre-Wisconsin drifts and border of Mankato(?) drift omitted. Drift borders dashed where approximately located.

B—Border of Binghamton drift (MacClintock and Apfel, 1944; MacClintock, 1954)

BB—Buzzards Bay moraine (Mather, Goldthwait, and Thiesmeyer, 1942)

BD—Southern limit of Binghamton drift in Elmira region (Denny, this paper)

C—Charlestown (Watch Hill) moraine (Woodworth, 1896a)

HH—Harbor Hill moraine (Woodworth, 1901; Veatch, 1903, 1906; Fuller, 1914)

LW—Late Wisconsin terminal(?) moraine stage (Rich, 1935)

N—Nantucket moraine (Woodworth and Wigglesworth, 1934)

OL—Border of Olean drift (MacClintock and Apfel, 1944; Leverett, 1934; Denny, in press)

R—Ronkonkoma moraine (Veatch, 1903; Fuller, 1914)

S—Sandwich moraine (Mather, Goldthwait, and Thiesmeyer, 1942)

VH—Valley Heads moraine (Fairchild, 1932; Williams, Tarr, and Kindle, 1909; Holmes, 1952)

If all the drift in the Elmira region south of Fairchild's Valley Heads moraine was deposited during Olean time, what is the explanation for the isolated deposits of kame gravel and till that contain numerous erratics? The following hypothesis is offered: Prior to Olean time the major valleys contained gravel with abundant erratics (i.e. gravel with "Binghamton" lithology)—perhaps early Wisconsin or pre-Wisconsin outwash.[2] During Olean time the ice sheet advanced from the northeast, roughly parallel to the striae and at right angles to the drift border. The Olean ice sheet picked up the older gravels and incorporated them in the adjacent drift. In this way the ice sheet deposited drift with numerous erratics on the sides of the major valleys, but by the time it reached the adjacent uplands most of these erratics had been left behind, and it quarried and deposited fragments of the local bedrock. Thus the Binghamton drift border of MacClintock and Apfel (fig. 1-B) where it closely follows the south side of the major east-trending valley of the Elmira region would mark the southern limit of Olean drift containing abundant erratics derived from a pre-Olean gravel fill in the adjacent valley. It is this lithologic change in the drift that was believed by MacClintock and Apfel to mark the border of Binghamton drift. Using their criteria the writer determined that the boundary lies as much as 10 miles south of the position located by them (fig. 1-B) in the area southwest of Elmira.

COMPARISON OF DRIFTS IN THE ELMIRA REGION WITH THOSE OF SOUTHERN NEW ENGLAND

The drift of the Valley Heads substage (Fairchild, 1932) and the drift of southern New England (Massachusetts, Rhode Island, Connecticut, and Long Island) have several features in common, such as similar topographic expression and abundance of water-laid material. On the other hand the Olean drift and the drift in southern New England differ in topographic expression, in relative volume of water-laid material, and in relative abundance of evidence of periglacial frost action. The cause of these differences is probably climatic: during Olean time and during advance of Valley Heads ice a rigorous climate very likely induced mass movements and formation of patterned ground; whereas during retreat of Valley Heads ice and the last ice over southern New England, a milder climate probably caused only slight modification of drift topography and only a limited development of patterned ground. If this is so, then these similarities and differences are valid criteria for correlation.

Valley Heads drift and drift of southern New England have many topographic features commonly found in other glaciated regions, whereas the Olean drift is unusual and is largely a smooth-surfaced mantle. Kame terraces are one of the most widespread topographic features in southern New England and in the Finger Lakes region north of the Valley Heads moraine. Kame terraces are scarce in areas covered by drift of the Olean substage.

The proportion of water-laid material in the Valley Heads drift and in the drift of southern New England is probably several times greater than that in the Olean drift. In the Elmira region and adjacent Pennsylvania, not more

[2] No pre-Olean deposits of Wisconsin age are known, but pre-Wisconsin outwash gravel occurs near Williamsport, Pa. (Leverett, 1934).

than about 5 percent of the Olean drift is water-laid, and near the border of the drift the amount is perhaps not more than 2 percent. In southern New England, on the other hand, water-laid drift covers perhaps from 10 to 20 percent of the land surface, and in addition, the Wisconsin drift on Long Island consists chiefly of water-laid material. This difference suggests that much smaller volumes of meltwater were produced during deglaciation of the Olean ice sheet than during deglaciation of the Valley Heads ice sheet or the last ice sheet in southern New England and Long Island.[3]

In the Elmira region and in adjacent Pennsylvania, patterned ground and rubbly colluvial deposits mantle the drift of the Olean substage (Peltier, 1949; Denny, 1951, in press; Smith, 1953). Blockfields, boulder rings, and boulder stripes are common features in areas underlain by the Pottsville and Pocono formations (Stose and Ljungstedt, 1931). Colluvial deposits, chiefly rubble, mantle the drift in many places and locally bury kame terraces. These colluvial materials and the patterned ground are periglacial deposits and soil structures resulting from cold climates during either the disappearance of the Olean ice sheet or the advance of the Valley Heads ice sheet or both. Within the area covered by drift of the Valley Heads substage and in southern New England, evidences of periglacial frost action are relatively scarce, and the features reported (Denny, 1936, 1938; Raup, 1951; Stout, 1952) are poorly expressed in comparison with those found south of the Valley Heads moraine.

These comparisons suggest that all the drift of Wisconsin age in southern New England and on Long Island is younger than the drift in the Elmira region and is perhaps to be correlated with Fairchild's Valley Heads drift.

POSSIBLE EASTERN EQUIVALENTS OF THE VALLEY HEADS MORAINE

East of the Finger Lakes region (fig. 1-C), this moraine has not been mapped in detail. Perhaps it roughly follows MacClintock's line (fig. 1-B) to join Rich's "late Wisconsin terminal(?) moraine stage" in the Catskill Mountains. Rich's description of the features on either side of his moraine is similar to that just given for the Valley Heads moraine south of the Finger Lakes.

[The moraine is] "the boundary between a region of prevailing fresh loop moraines, strong glacial scouring and fresh, little eroded drift [to the north and east] . . . and a region [to the south and west] . . . where the characteristic features are smooth till or thick drift of subdued form, few distinct loop moraines, noticeably greater erosion of the drift by postglacial streams, large postglacial alluvial fans and in places weathered boulders in the drift" (Rich, 1935, p. 120-121).

South of the Catskill Mountains a possible equivalent of the Valley Heads moraine is unknown but may perhaps follow the western side of the Delaware Valley to the terminal moraine of the Wisconsin stage (fig. 1-C). Eastward the equivalent is probably the terminal moraine in New Jersey. On Long Island and to the east (fig. 1-C) the limit of the Valley Heads ice was prob-

[3] It is possible that the relative abundance of stratified drift in southern New England may result partly from an abundance of coarse sizes in the ice load (J. H. Hartshorn and J. P. Schafer, written communication, Feb. 1, 1955).

ably the Harbor Hill, Charlestown,[4] Buzzards Bay, and Sandwich moraines (Mather, Goldthwait, and Thiesmeyer, 1942), although there is no evidence to exclude the possibility that the Valley Heads moraine of Fairchild (1932) might be equivalent to the Ronkonkoma and Nantucket moraines.

The Valley Heads moraine and its probable equivalent in the Catskill Mountains are prominent topographic features, and logically their eastern equivalents should be comparable. In southeastern New York State and in southern New England the only topographically similar moraines are on Long Island and Cape Cod or the islands to the south. The moraines recognized in central New England prior to about 1920 (Taylor, 1903; Alden, 1924) are now attributed to ice stagnation (Flint, 1930, 1933; Goldthwait, 1938; Currier, 1941a; Jahns, 1953). Flint (fig. 1-A) and MacClintock (fig. 1-B) placed a tentative border of Cary drift across southern New England where there are few remnants comparable in size and continuity to those in New York State except those near Middletown, Connecticut, discussed below. Why should the Valley Heads ice have built a prominent moraine in the Finger Lakes region and across the Catskill Mountains but have left no comparable belt of moraine across southern New England? East of the Hudson River, prominent moraines are found chiefly on Long Island and Cape Cod.

FLINT'S CORRELATION

Flint (1953) tentatively placed the border of the Cary drift across southern New England on a line that runs from Middletown to Boston (fig. 1-A). The principal evidence is (1) superposition of drifts, (2) weathering of older drift (Iowan-Tazewell complex) where buried by younger drift (Cary?), (3) ventifacts and evidence of periglacial frost action on drift of the Iowan-Tazewell complex and their scarcity on Cary(?) drift.

The strongest evidence for Flint's suggested Cary(?) drift border (fig. 1-A) is in and near the Connecticut River Valley near Middletown. Drift overlies disturbed varved sediments throughout an east-trending belt more than 16 miles wide (Flint, 1953). North of Middletown a great mass of pitted outwash, now trenched by the Connecticut River, dammed the valley to form a lake recorded by another and separate body of varved sediments that extends northward for more than 50 miles (Flint, 1933; Jahns and Willard, 1942). The lake drained southward through a gap near New Britain, about 10 miles northwest of Middletown. Thus a readvance of at least 16 miles followed by a prolonged retreat is demonstrable in the Middletown region.

To the east, however, the location of the Cary drift border is indefinite. A loose, sandy till overlying a brownish, slightly indurated and compact till has been described in northeastern Connecticut by White (1947) and in northeastern Massachusetts by Currier (1941b), Judson (1949), and others. The southern limit of this "younger" till is unknown, but Flint noted (1953, pl. 1) that all these localities lie north of the southern limit of his Cary(?) drift. However Chute (1949, 1950) described superimposed drifts, from localities

[4] The Charlestown moraine (Woodworth, 1896b, p. 152; 1896a, table opposite p. 988; 1898, fig. 7; Woodworth and Marbut, 1896, fig. 2; Woodworth and Wigglesworth, 1934, table, p. 37) has recently been called the Watch Hill moraine (Bryan and Ray, 1940, p. 61-62; Antevs, 1945, p. 4). The term "Charlestown" clearly has priority.

south of the Middletown-Boston line, that he believed were deposited by one ice sheet. White (1947) held that his "older" till was subaerially weathered prior to deposition of his "younger" till, and this was cited by Flint (1953) as evidence of a break of substage rank between his Iowan-Tazewell complex and the drift of Cary(?) subage. However, it has never been firmly established that the features described by Currier (1941b), White (1947), Chute (1949, 1950), or Judson (1949) are actually the result of subaerial weathering. No organic matter is reported. If the weathered(?) horizons are parts of soil profiles, probably they are remnants of the B or C horizons, or both. Most descriptions mention only that the till is stained by limonite or is brown and more compact than the material above. Such limonite staining of drift is perhaps related to iron sulfides therein, or to weathering by ground water. The occurrence of limonite staining and compaction is not necessarily evidence that the drift was modified by soil-forming processes.

Ventifacts in southern Connecticut and in the Cape Cod region were mentioned by Flint (1953, p. 900) as indicative of "the glacial-maximum conditions evident at the Iowan and Tazewell drift borders in Iowa and in South Dakota." However, ventifacts are numerous also in eastern Massachusetts north of the Middletown-Boston line; they also occur in New Hampshire (Goldthwait et al., 1951). On the other hand, ventifacts are essentially unknown along the border of MacClintock and Apfel's Binghamton drift (and also along the border of the Valley Heads drift) in southern New York State, which Flint tentatively correlated with the Cary substage (fig. 1-A).

Disturbance of drift of the Iowan-Tazewell complex by periglacial frost action is noted as a common feature in southern Connecticut (Flint, 1953, p. 899). However, such disturbance is also reported north of the Middletown-Boston line (Raup, 1951; Stout, 1952), and the deposits and structures resulting from periglacial frost action are much thicker and more abundant in areas of Olean drift than anywhere in southern New England or on Long Island.

Flint (1953, p. 901) stated that the Iowan-Tazewell complex "is continuous with the outer Wisconsin drift from Long Island westward across the Appalachian region into the Central Lowland, where it is known to include the Tazewell substage." However, the two and possibly three (Fleming, 1935) advances present near the margin of the complex in southern New England have not been recognized in the Olean drift of northeastern Pennsylvania. If my correlation of Fairchild's Valley Heads moraine with the Harbor Hill moraine and associates is correct (fig. 1-C), MacClintock and Apfel's Binghamton drift logically correlates with the Ronkonkoma and Nantucket moraines. MacClintock's (1954) belief that the Binghamton drift is closely related in time to the Valley Heads drift (both of the Cary substage, fig. 1-B) is probably correct except for the doubts already expressed of the existence of Binghamton drift in the Elmira region.

Flint (1953) discussed the relation between Lake Hackensack—a glacial lake that extended from the terminal moraine of Wisconsin age along the northeast border of New Jersey and north nearly to the border of the Cary drift (fig. 1-A)—and Lake Albany, which extended from the Cary border

northward along the east side of the Catskill Mountains. He cited convincing evidence that the two lakes were separate and distinct and "suggest the presence of a glacier terminus south of Lake Albany, but north of Lake Hackensack." However, the supposition that, following development of Lake Hackensack, the ice edge retreated northward (during the Tazewell-Cary interval) and then readvanced to the Cary drift border, while logical, is not supported by stratigraphic evidence, such as is present north of Middletown. Could not the hypothetical drift dam at the south end of Lake Albany have been a recessional feature built during retreat of ice from the Wisconsin terminus? Antevs' study of varves (Antevs, 1928) supports Flint's hypothesis; the deposits of Lake Albany are correlated with those of the Connecticut Valley Lake (Cary substage) and those of Lake Hackensack are correlated with lake deposits in the Iowan-Tazewell complex (in the lower Quinnipac Valley).

Thus the readvance of an ice sheet for at least 16 miles is established in the Middletown region, but the supposition that there was a major ice retreat prior to this advance, during the Tazewell-Cary interval, rests on tenuous evidence. The existence of a drift border from Middletown to Boston or westward into southeastern New York State is also based largely on assumptions. I submit that most of the evidence cited by Flint (1953) for a possible Cary subage of the drift north of the Middletown-Boston line is also applicable to the drift to the south. If the proposed correlation of Fairchild's Valley Heads moraine with those on Long Island and Cape Cod is correct, this in itself does not eliminate the possible existence of other drift borders across southern New England.

MACCLINTOCK'S CORRELATION

MacClintock's (1954) interpretation of the drift borders of the Wisconsin stage in parts of northeastern United States (fig. 1-B) differs from the interpretation presented in this paper chiefly in two respects. First, as already discussed, MacClintock recognized Binghamton drift in the Elmira region; and second, in northern New Jersey and vicinity, MacClintock recognized two Wisconsin drifts (Tazewell and Cary) where I identify only one drift (that of the Valley Heads substage). However, MacClintock's data can be interpreted to support my interpretation.

MacClintock (1954) studied the depth of leaching of calcareous outwash gravel as a possible indicator of age of drift. He determined the depth of leaching of a gravel deposit by testing its sandy matrix with dilute HCl (table 1,[5] column II). The amount of carbonate material originally present in the gravel was measured by making pebble counts of the unweathered calcareous material below depth of leaching (table 1, columns III to VII incl.). Note that in table 1, column V equals the sum of columns III and IV. Table 1 shows that MacClintock's Tazewell drift (column IX) ranges from 7 to 16 percent pebbles of calcareous rock and from 11 to 15 feet in depth of leaching; his Cary drift ranges from 11 to 49 percent pebbles of calcareous rock, and from 4½ to 8 feet in depth of leaching.

[5] Table 1 presents MacClintock's data for localities in northern New Jersey and vicinity (MacClintock, 1954, table 1).

It is here suggested that the data in table 1 do not demonstrate conclusively that the gravel deposits belong to two substages. It is not possible that the differences in depth of leaching of these gravels are related chiefly to the total amount of free carbonates in the gravels rather than to their content of calcareous pebbles or to the length of time that they have been undergoing leaching? In northwestern New Jersey (fig. 1-B) MacClintock's Cary drift is underlain in part by the Cambrian and Ordovician "Kittatinny" limestone and other limestones of Silurian and Devonian age, and his Tazewell drift rests chiefly on Precambrian crystalline rocks and on slightly calcareous or noncalcareous sedimentary and igneous rocks of Triassic age (Lewis and Kümmel, 1912). If it is assumed that the gravels are chiefly of local derivation, it follows that MacClintock's gravels of the Cary substage are composed predominantly of limestone pebbles (columns III and IV, table 1), whereas his Tazewell gravels are predominantly slightly calcareous and noncalcareous rock (column V, table 1).

TABLE 1

Depth of Leaching and Pebble Content of Gravels of Wisconsin Age in Northern New Jersey and Vicinity (MacClintock, 1954, fig. 5 and table 1)

		Percent of pebbles						
I	II	III	IV	V	VI	VII	VIII	IX
Station	Depth of leaching (feet)	Lime-stone	Dolomite	Total Car-bonates	Sandstone and siltstone	All others	Total stones	Substage T = Tazewell C = Cary
1	11			7	89	3	1168	T
2	12-15			9	1	90	217	T
3	11-12			16	16	66	198	T
4	11			11	12	77	150	T
6	14-15			11	7	81	230	T
25	6-7	15	3	18*	44	37	199	C
28	7	7	4	11*	73	16	155	C
29	7	11	2	13*	50	38	338	C
30	6			49	24	27	100	C
31	4½	12	23	45**	20	43	208	C
43	4-5	5	33	38*	20	42	131	C
114	7-8	7	4	11*	73	16	155	C
115	5-8	15	3	18*	44	37	199	C
116†	5-8	11	2	13*	36	2	338	C

* Percentage of limestone and dolomite.
** Percentage probably incorrect, should be 35.
† Total percentage 51.

MacClintock (1954) divided the calcareous pebbles into 3 groups—limestone, dolomite, and total carbonates (table 1). The deeply leached gravels (Tazewell drift) contain no pebbles of limestone or dolomite, but from 7 to 16 percent of the pebbles are calcareous, presumably slightly calcareous sandstone and siltstone. Almost all the shallowy leached gravels (Cary drift) contain from 11 to 45 percent of pebbles of limestone and dolomite. If the

free carbonate content of the matrix of the deeply leached gravels is largely derived from adjacent outcrops of slightly calcareous bedrock, then perhaps the total content of free carbonates (matrix plus pebbles) is considerably less than that of the shallowly leached gravels. The latter may have a higher total content of free carbonates because their matrix is doubtless chiefly derived from adjacent outcrops of limestone and dolomite. Thus the depth of leaching of these outwash gravels in New Jersey may be related in large part to their total content of free carbonates and not to the length of time since deposition. Under my interpretation all these gravels belong to the Valley Heads substage.

CONCLUSIONS

In the Elmira region the drift of the Olean substage contrasts topographically both with the younger Valley Heads drift of Fairchild and the drift of southern New England. Similarly water-laid drift of the Olean substage is relatively scarce compared with that in the Valley Heads drift and in southern New England. In addition, evidence of periglacial frost action is common outside of the Valley Heads moraine and is scarce to the north and in southern New England. These three facts suggest that drift of the Olean substage is not present at the surface in southern New England and Long Island, and that the Valley Heads drift of Fairchild is perhaps the same age as the drift of southern New England and Long Island. Thus the Valley Heads moraine of Fairchild may be equivalent to Rich's (1935) "late Wisconsin terminal(?) moraine stage" in the Catskill Mountains and to the Harbor Hill, Charlestown, Buzzards Bay, and Sandwich moraines (fig. 1-C). The Binghamton drift of MacClintock and Apfel (1944) is not present in the Elmira region (fig. 1-C) where it apparently was buried by the Valley Heads ice sheet. However, Binghamton drift probably is present east of the Elmira region (fig. 1-C) between it and the Catskill Mountains. Binghamton drift was probably deposited just prior to the building of the Valley Heads moraine, and this drift is possibly equivalent to the Ronkonkoma and Nantucket moraines of Long Island and southern New England (fig. 1-C).

In the Elmira region the topographic distinction between the Olean drift on the one hand and Valley Heads drift on the other is difficult to explain. Is it an initial depositional feature or was the Olean drift "smoothed" by mass movements during advance of the Valley Heads ice sheet? The rubble-mantled kame terraces and other evidence suggest that the Olean drift was modified by mass movements prior to building of Valley Heads moraine—indicating a time interval between formation of some Olean kame terraces and deposition of their mantle of rubble. The length of this interval is unknown. The difference in volume of water-laid drift suggests differences in mode of deglaciation of the Olean ice sheet and of the Valley Heads ice sheet, but other explanations of this difference are permissible. In the absence of stratigraphic evidence it is impossible to say whether there was an interval of prolonged ice retreat between the deposition of the Olean drift and the deposition of the Valley Heads drift or whether, as Tarr believed (Williams, Tarr, and Kindle, 1909), the Valley Heads is a recessional moraine built by

a minor readvance of the ice sheet during its retreat from the terminal moraine (Olean drift border, fig. 1-C) of the Wisconsin stage (see also Von Engeln, 1921, 1929).

The correlation of the drifts in the Elmira region with those in the Central Lowland rests on exceedingly tenuous evidence. Ruhe (1952a, 1952b) has suggested a two-fold division of the Wisconsin stage based in part on topographic evidence. It is interesting but far from conclusive that the topographic contrast between his lower and upper Wisconsin drifts is reminiscent of that between the Olean drift and the Valley Heads drift of the Elmira region. If this correlation is possible, then perhaps his Bradyan interval is the time interval of unknown duration between the Olean and Valley Heads drifts of the Elmira region.

The reconstruction presented in this paper places the southern limit of ice during Valley Heads time considerably farther south in New England than in south-central New York State. The explanation might involve topographic differences between the two areas. Possibly the explanation is related to local centers of ice accumulation in northern New Hampshire and Vermont (Flint, 1951) or to climatic effects induced by the adjacent ocean.

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LATE PLEISTOCENE FEATURES OF CHEBOYGAN AND EMMET COUNTIES, MICHIGAN

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ABSTRACT. Detailed mapping of the surficial glacial and postglacial landscape features in Cheboygan and Emmet counties at the northern tip of the southern Michigan peninsula confirms that practically all the area was glaciated by Valders ice moving in from the northwest across the Lake Michigan basin. Only a thin and weak ice mass overrode the Emmet moraine with the result that the major ice advance was stabilized along a front just inland from the present shores of Lake Michigan and the advance sheet stagnated in situ over the two-county area. The present topography is largely the result of the distribution of meltwater channels, ice blocks, and outwash plains. These have been much modified by three succeeding lake strandlines, (1) the pre-Algonquin level here identified for the first time, and probably representing Toleston Lake, (2) the slightly higher Algonquin level, and (3) the Nipissing level. The wet-climate Algonquin level is separated in time by a long period coincident with Lake Chippewa before the occurrence of the dry-climate Nipissing level.

INTRODUCTION

The tongue of land dividing Lakes Michigan and Huron on the south contains a complex of lakes, old shore features, and hills. Its postglacial aspects invite study in the light of recent additions to our knowledge of the late Pleistocene.

Leverett and Taylor (1915) mapped the hills of Cheboygan and Emmet counties as an interlobate morainic area between the Huron and Michigan lobes of what has come to be known as the Cary substage of Wisconsin glaciation. Their reconnaissance, supplemented by detailed descriptions of Scott (1921) established the locations, elevations, and basic correlations of the old shorelines in the area.

Although the existence of a red till in the area was remarked by Leverett and Taylor—and was tentatively correlated with a similar till on the Wisconsin side of Lake Michigan north of Milwaukee—these tills were assumed to mark merely a slight readvance from the northeast, the direction of Cary ice movement. In recent years, the red till in Wisconsin has been found to result from a major ice advance from the northwest. This advance, the Valders substage of the Wisconsin (Thwaites, 1943), may possibly be contemporaneous with the Mankato substage in Minnesota and is dated by the fact that it overlies the Two Creeks forest bed in Wisconsin, the age of which has been fixed at about 11,400 years by four radiocarbon laboratories.

The object of the present paper is to present the details of this last ice advance in lower Michigan as well as the features of withdrawal and of the subsequent stages of the Great Lakes.

METHODS

A detailed reconnaissance of the area surrounding the University of Michigan Biological Station was undertaken by the senior author during the summer of 1954. This reconnaissance was extended to cover virtually all of Cheboygan and Emmet counties although the southeastern portion of Cheboygan County was covered only sketchily.

* A contribution from the School of Natural Resources and the Department of Geology, University of Michigan, based on field investigations at the University of Michigan Biological Station in the center of the study area.

Surface geological features of the area were mapped under stereoscopic observation using the 1952 coverage of the Production and Marketing Administration, U. S. Department of Agriculture, and the Michigan Department of Conservation. These modified infrared photographs were at a scale of 1:15,840 or 4 inches to the mile. All important and many minor features were visited and checked in the field. Several hundred elevations were determined by Paulin System surveying barometers, using rapid closed traverses from U. S. Coast and Geodetic Survey benchmarks or from the shores of Lakes Michigan and Huron. The convergence of several railroads and highways at the Straits of Mackinac, coupled with the existence of good vertical control data along these thoroughfares, made the area well suited for this type of reconnaisance survey.

VALDERS GLACIATION

For many years, the realization has grown that the last major ice advance to reach the lower peninsula of Michigan is represented by a red till in the upper part of the peninsula. Leverett and Taylor (1915) correlated red till from the Port Huron moraine with red till in Wisconsin from Milwaukee north that had been previously studied by Alden. By 1943, Thwaites had gathered data indicating that these tills represented a major readvance, a theory confirmed for the Michigan deposits by Bretz (1951). More recently, Melhorn (1954) has worked out the details of the Valders glaciation in northern lower Michigan. Most of Cheboygan and Emmet counties lies within its borders (fig. 1).

In these counties, the southeasterly direction of ice movement is everywhere apparent in drumlins (Leverett and Taylor, 1915; Bergquist, 1943) and in ridging and channelling of the ground moraine. The more prominent drumlins in the two countries are mapped in figure 1. The same southeasterly direction of ice movement is visible in the ground moraine.

Valders deposits tend to be red in color (Munsell 5YR, 6-7/2-3), finer-grained than comparable Cary deposits, and somewhat higher in lime content. The red coloration is apparently derived from deposits in the Lake Michigan basin which themselves were derived from materials in the Lake Superior area. Furthermore, the sandier the soil, the lighter the color inasmuch as the color is in the silts and clays and only to a minor extent in the larger particles. The carbonate content is derived locally from limestone bedrock and seems to decrease as these areas are left behind. The finer particle size confirms the findings of Murray (1953) and Melhorn (1954).

The Valders ice sheet was apparently thin and weak. Its energy was largely dissipated in overriding the high land in northern Emmet County and it formed few constructional features southeast of this area. Furthermore the ice apparently stagnated in situ over this broad belt so that terminal and recessional moraines were not developed. The largest and longest-lasting ice blocks gave rise to Douglas Lake, Burt Lake, Black Lake, and numerous smaller lakes and swamps, the most important of which are indicated on figure 1. The largest ice-block extended some 20 miles from near Cheboygan to the far end of Black Lake. Well developed ice-contact faces remain on its

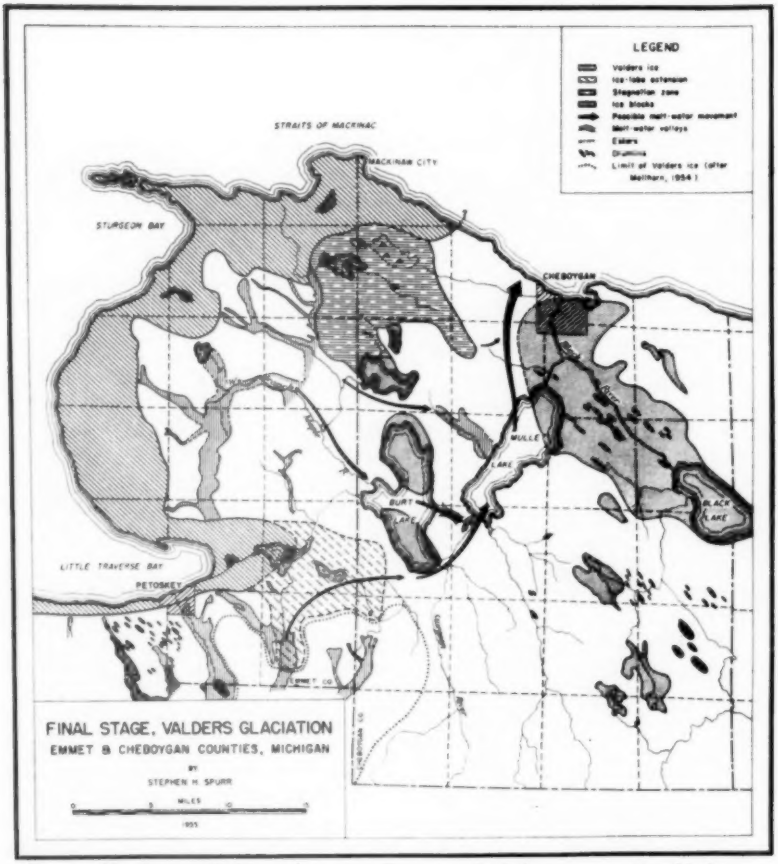

Fig. 1. Final stage, Valders glaciation, Emmet and Cheboygan counties, Michigan.

north side, and it covered a drumlin area sufficiently long so that only a thin veneer of sands was deposited under the subsequent pre-Algonquin and Algonquin Lakes. Its western limit is apparently marked by an underwater ridge which nearly reaches the surface in the northern portion of Mullet Lake.

The most spectacular ice-melt features in the area are the several large, entrenched, and clearly defined meltwater valleys. The first one to be found and studied in detail is now largely occupied by Van Creek. A pronounced valley nearly three-quarters of a mile wide with a stream channel nearly one-quarter of a mile wide emerges from the Algonquin Island south of Levering. The channel may be traced on aerial photographs three miles to the southeast even though it has been since planed by Lake Algonquin. Upstream, a well-defined valley oriented with the grooved Valders ground mor-

aine is easily traced to the foot of high kames lying at the head of an ice-tongue depression. The ice-contact slopes surrounding this depression stand 180 feet above its floor and the head of the meltwater valley lies at 835 feet above sea level or 70 feet above the depression. Clearly, here stood the live ice mass during the stagnation of the thin ice to the southeast.

Once the feature was recognized, similar situations became apparent throughout the area. The more prominent meltwater streams are mapped in figure 1. Most of these are obliterated as they pass beneath the level of Lake Algonquin. The meltwater channel between Burt and Mullet lakes, however, survived due to its protected location.

In the stereoscopic view of the aerial photographs, it is possible to see with fair precision the line of ice-contact slopes connecting the heads of these meltwater channels and marking a relatively stable position of the Valders ice shortly after its maximum advance. Two stages of the Little Traverse Bay ice tongue, though, need to be postulated on the basis of drainage features (fig. 1). The maximum stage drained southward through the meltwater channels at the present Emmet and Cheboygan county boundaries, while the second stage drained eastward and northward around the Burt Lake ice-block, and down the channel now occupied by Mullet Lake.

The northern finger lakes such as the two arms of Walloon Lake occupy south-flowing meltwater channels which drained the Little Traverse ice lobe south and west into the proglacial lake occupying the Michigan basin south of the Valders ice. The present Bear Creek valley was originally such a south-flowing meltwater channel, but one that has subsequently been captured by a small stream cutting headward a few hundred yards from the later Algonquin shore.

Since Lake Algonquin waters have covered and obliterated most of the lower meltwater features, any effort to reconstruct the pattern of meltwater movement must remain hypothetical. However, the blocking of Little Traverse Bay by an arm of the Valders ice necessitates drainage to the Huron Basin, very likely around the west end of the Black Lake ice-block. Furthermore, it appears logical to connect the Van Creek valley with the group of small kettles south of Douglas Lake and the older valley of Nigger Creek to provide one major drainage channel. Similarly, the West Maple meltwater channel is continuous in a direction parallel with Valders ice movement with the lower Maple River channel, the irregularity in Burt Lake indicating a gap in the ice-blocks, and the channel between Burt and Mullet Lakes. Both streams, as well as the drainage from the Little Traverse lobe would have moved up the channel now occupied by Mullet Lake to the Huron basin. In such a situation, the capture of the Van Creek stream by the Maple meltwater stream west of Douglas Lake would account for the subsequent filling of that portion of the Van Creek channel immediately south of the Douglas Lake ice blocks, and the development of small kettles in the abandoned channel.

No evidence on the ground was found to substantiate the "Cheboygan moraine" of Leverett. Rather, elements of a northwest-southeast-trending ridge between Mackinaw City and Cheboygan seem to be made up of Valders

till trending in the same direction, underlain by limestone bedrock, and accentuated by the Nipissing wave-cut cliff.

PRE-ALGONQUIN LAKE

As the ice melted, the waters of the proglacial lake rose, eventually reaching the strongly defined Algonquin shore. Beginning with Leverett, various writers have assumed that the Algonquin level was the first major lake level after the retreat of the ice from the Michigan basin, and that the slightly lower strandlines throughout the Emmet-Cheboygan area represent a recessional stage from the Algonquin high.

As the present reconnaissance progressed, however, more and more doubt was raised about this recessional stage. First, some of the supposed Algonquin

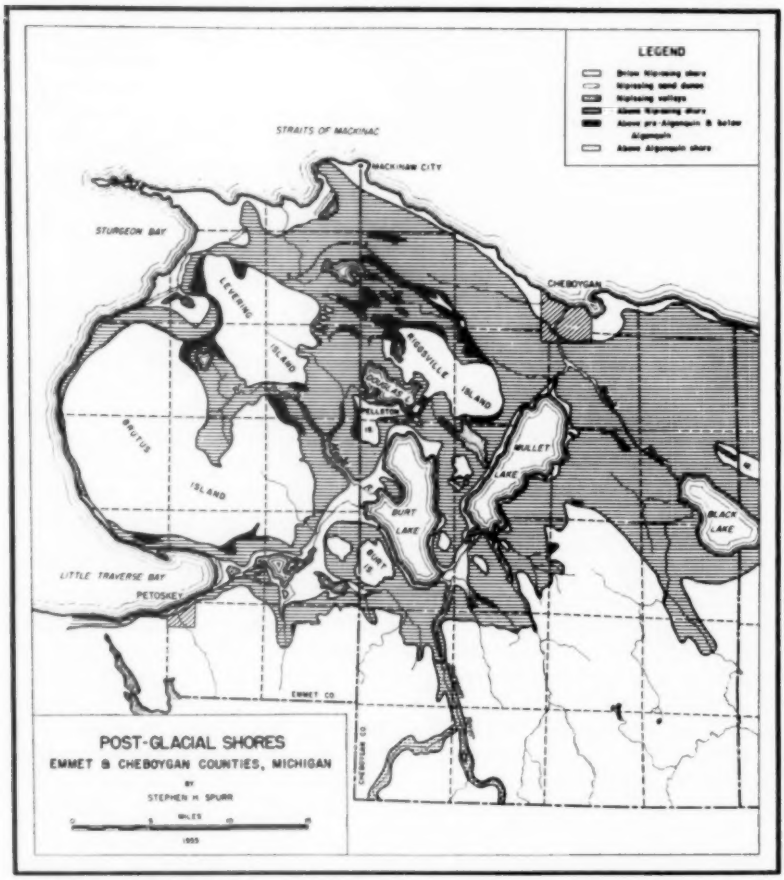

Fig. 2. Postglacial shores, Emmet and Cheboygan counties, Michigan. Sturgeon River valley is of Algonquin age.

shore features were found to be 18-25 feet lower than equally well-defined shore features nearby. Then, cases were turned up of two well-defined shores, one above the other. Gradually, it became apparent that there were two major lake levels over 700 feet above sea level, and that the lower one was the older. In sheltered locations such as along the north edge of Little Traverse bay, the pre-Algonquin shore is clear-cut except where it approaches the higher Algonquin shore and is obliterated by the wave-worked sands of the younger level. On windward shores, the earlier shoreline is completely obliterated; but in many instances the Algonquin lake merely held the same shoreline as indicated by the lower elevation of the combined wave-cut terrace.

Once a series of clear-cut examples of the two shores had been located at various points in the two counties, it became possible to recognize the pre-Algonquin shore throughout the area. Its major features are indicated in figure 2. Actually, this map is incomplete in so far as the pre-Algonquin shore is concerned because its existence was only fully demonstrated towards the end of the field season, and because the features are so largely obliterated by the higher Algonquin level in many areas that it was not felt worthwhile to fill in all the detail. Very probably, the Algonquin Islands named Brutus and Levering formed a single land mass in the pre-Algonquin lake, joined by kamic deposits north of the meltwater valley the head of which is now filled by Larks Lake. Less clearly, a partially obliterated but still visible shoreline northwest of Carp Lake indicates that this land mass continued easterly to include Riggsville Island. In Algonquin times, the intervening area, built largely of outwash sands, was planed to create a shallow lake bottom dotted with small residual islands aligned with the direction of Valders ice flow.

The pre-Algonquin lake represents an early stage of Lake Algonquin, and was formed after the ice had retreated sufficiently from the Michigan basin to allow the merging of Michigan and Huron basin waters through the Mackinac channel. It is named "pre-Algonquin" here to distinguish it from the commonly recognized Algonquin shore of the last stage of Lake Algonquin. It is consistently 18-25 feet lower than the Algonquin shore. Whether the bend in the pre-Algonquin shore shown in figure 4 is real or merely the reflection of errors in the aneroid determinations of elevations can only be determined by spirit-level work. The aneroid elevation data are plotted as they were obtained without any effort to interpret the scatter into a set pattern.

The pre-Algonquin shore varies in elevation from 684 feet at Harbor Springs where it is exposed for several miles in two sections to about 730 feet along the north edge of the morainic-outwash land mass both east and west of Carp Lake. Its interpolated elevations at Mackinaw City, the University of Michigan Biological Station, and Petoskey are given in table 1. It is apparently represented by the strongest of the "upper group" of Algonquin strandlines on Mackinac Island (Leverett and Taylor, 1915, p. 425).

LAKE ALGONQUIN

The Lake Algonquin shore is almost everywhere well defined and prominent. Throughout the two counties, it is represented by a wave-cut terrace

and cliff in till or outwash deposits. Only south of Cross Village is a constructional beach developed over a two-mile stretch.

In Lake Algonquin, the lower peninsula of Michigan terminated south of Little Traverse Bay and the area to the north contained a complex archipelago. Lake waters widened the meltwater channel now occupied by the West Maple River and cut across the deposits north of Larks Lake to separate Brutus Island from Levering Island. Similarly, the till deposits on Riggsville Island were separated from the outwash sands of Pellston Island by two channels cut near the present Biological Station. The details of Lake Algonquin given in figure 2 follow in their broad outlines earlier mapping by Leverett and Taylor (1915) and others. On figure 2, the dashed lines in the Sturgeon River Valley demarcate the Algonquin river terraces.

The Algonquin shores everywhere exhibit a characteristic pattern, most easily seen on the aerial photographs. The shore features are stable, mature, and strongly marked by many rills and minor water courses. Sand dunes are rare and minor. The tributary streams are broad and are perfectly graded to the level of the lake. Furthermore, there were many more active streams than at the present time, and the channels of the surviving streams were considerably larger than at the present time. With experience, the Algonquin shore could be recognized by its geomorphology without prior determination of its elevation. Clearly, Lake Algonquin was a stable lake existing in a climate substantially wetter than that of the present time.

TABLE 1

Strandlines in the Cheboygan-Emmet Area

Name of strandline	Present elevation above sea level			Features
	Mackinaw City	U.M.B.S.	Petoskey	
Lake Michigan Lake Huron	581	581	581	Modern shoreline
Algoma	618?	592?	—	Faint beaches along Lake Huron
Nipissing	630	620	612	Well developed
Chippewa	low—perhaps 500			Inferred from upper valleys of Maple River and Nigger Creek
Algonquin	780	738	698	Well developed
Pre-algonquin	750	724	672	Residual shores in sheltered areas
Local ice-melt	—	circa 630-660	—	Indicated by gradient of Pellston Plain

The pink clay deposits underlying Douglas Lake and described by Wilson (1945) as originating under Lake Algonquin waters undoubtedly did originate under this lake, but probably at the earlier stage (here named pre-Algonquin) when the ice was closer to the Douglas Lake area and the Valders till was being actively reworked by the high lake waters. Parenthetically, the brown clay deposits described by Wilson may be Cary in age, thus antedating the Valders till.

On sheltered shores—largely those facing the southeast—numerous recessional strands of the Algonquin Lake can be discerned. An example is

the embayment at the north end of Burt Lake above the Nipissing shore. These are of seemingly little importance geologically and can usually be distinguished from the pre-Algonquin shore which lies near their upper margin.

LAKE CHIPPEWA

The working out of the deep-water river channel through the Straits of Mackinac by Stanley in 1938, the detection of shallow-water deposits deep in Lake Michigan by Hough in 1953, and the analysis of a buried soil profile near South Haven by Zumberge and Potzger (1955) have established the existence of a pronounced low-water Lake Chippewa between the times of Lake Algonquin and Lake Nipissing.

The present paper, dealing with features above the present lake level, cannot add much to this concept. At least two stream channels, however, seem to be graded to the Chippewa level. Nigger Creek is graded to an obviously low level and has been dammed and flooded by the waters of Mullet Lake

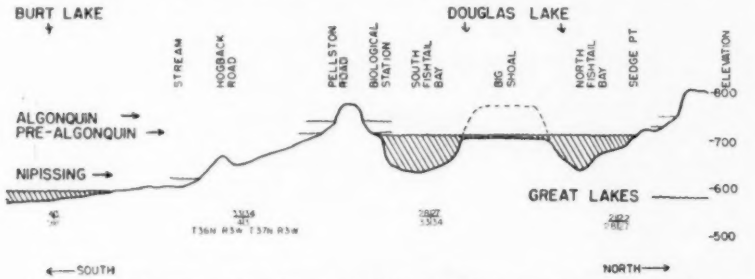

Fig. 3. North-south section through Biological Station. Based on topographic map prepared by former Engineering Camp, hydrographic charts of the U.S.C. and G.S. and Wilson (1945), and original surveys.

at 594 feet, only 13 feet above Lake Huron. A more pronounced feature, though, is an old valley edge in the Pellston sand plain due east of Pellston village. Where this edge lies above the Algonquin, it is represented by a constructional beach and dune area. Where it is cut into the Algonquin to the south, it is represented by a sharp valley edge. The valley edge disappears when it merges with the lower Nipissing valley and thus is older than that valley. Since this feature is clearly younger than Algonquin and older than Nipissing, and since it is graded to a lake level substantially lower than the present level (fig. 4), it may be considered a Chippewa feature.

LAKE NIPISSING

Like the Algonquin shore, the Nipissing shore is everywhere clear and prominent. Unlike the Algonquin shore, however, the geomorphology of the Nipissing shore is characteristic of a semi-arid climate. The wave-cut cliffs are steep and immature. The stream valleys are narrow, few and far between, and steep-walled. All the high large sand dune areas on the Michigan shore in Emmet County are apparently of Nipissing age. Many small dunes, most of which have since been stabilized, are found just above the Nipissing shores and near Nipissing valleys. The dunes south of the Biological Station sur-

mounted by Hogback Road (fig. 3) are examples of the shore dune features, while the dune area south of Pellston is clearly derived from the Nipissing channel of Maple River. As with the Algonquin shore, the Nipissing shore can be identified by its characteristic features, in this case those of a semi-arid climate. There seems but little doubt but that the equivalence of Nipissing and Xerothermic times postulated by Zumberge and Potzger is substantiated in the Cheboygan- Emmet County area.

The Nipissing time is of particular interest in the history of the inland lake chain. Mullet Lake began its postglacial history as a meltwater channel draining the stagnating Valder ice off to Lake Huron. As the ice retreated from Little Traverse Bay, the flow was reversed and a stream was formed that drained the Nigger Creek and Mullet channels down Indian River to the south Burt Lake ice-block lake, thence downstream to Crooked Lake and thence downstream to Little Traverse Bay. This stream continued through Chippewa times and became graded to that low level. As the lake water rose in Nipissing times, and reoccupied Little Traverse Bay, a row of sand dunes was constructed across the head of the bay, creating Round Lake. For a time, the inland lakes continued to drain around the south end of the dunes. At this time, the stream valley was flooded by the advancing Nipissing waters and a Nipissing Beach is clearly discernible around the Crooked-Burt-Mullet Lake chain (fig. 2). This beach surrounds the north end of Mullet Lake, however, and there was no Nipissing connection between Mullet Lake and Cheboygan Bay of Lake Huron. At Cheboygan, the advancing Nipissing

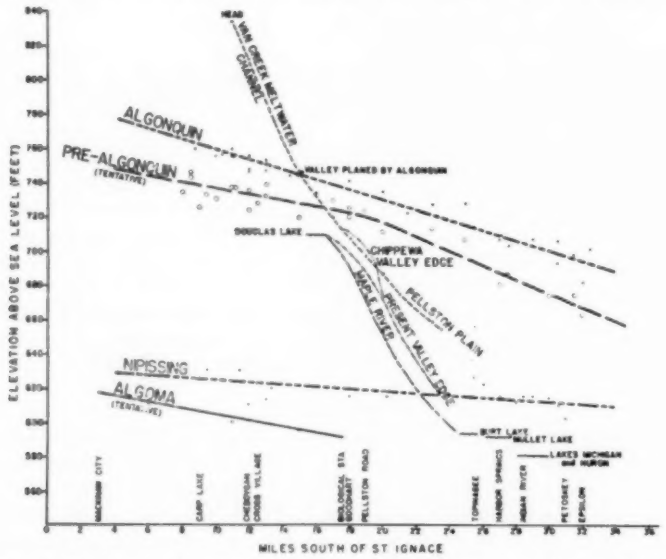

Fig. 4. Profiles of shorelines in Emmet and Cheboygan counties, and profiles of Pellston plain features. Intersection of Van Creek meltwater channel and Algonquin shore indicates point to which valley has been placed by Lake Algonquin.

waters again cut off a small stream which diverted its waters into Black River, a stream strong enough to maintain its channel through the offshore deposits.

In late Nipissing time, however, a tributary of the Black River worked headward less than a mile to the Nipissing shore of Mullet Lake and captured the inland lake area. These lakes persist to this day at a level controlled (until dam construction) by the gradient of this stream, now the upstream portion of the Cheboygan River.

Dune formation before the advancing Nipissing waters was most pronounced on the windward shores from Cross Village to Mackinaw City. Wycamp Lake, O'Neal Lake, and French Lake were all isolated from Lake Michigan at this time. Throughout this area, the high, partially stabilized Nipissing dunes are readily distinguishable from the lower, flatter dunes to the lakeward active at the present time.

As the Nipissing waters retreated, considerable dry land was created in the Wilderness Park area, Cheboygan Bay, and elsewhere. Many constructional recessional shores are evident but seem to be of little importance. The so-called Algoma beach is discernible at several points along Lake Huron, and its elevation is recorded in table 1. It is rather weakly developed, though, and it seems scarcely justifiable to classify it as representing a major lake stage.

It is of interest to note that practically all the white cedar (*Thuja occidentalis* L.) occurs below the Nipissing shore where leaching of the limestone from the surface soils has presumably been minimal. This may be seen around the north end of Burt Lake and elsewhere throughout the area.

SOILS

In a land surface of recent origin such as the Cheboygan-Emmet area, the soil types are closely related both to parent material and to drainage. A two-way classification based on these factors provides the best key to an understanding of the soil types.

The soils of the Biological Station Forest were described by Donahue (1936). Without using standard series names, he distinguished between pine lands, hardwood lands, conifer-hardwood lands, and swamp. Geologically, the distinction between these areas is clear. The pine lands consist of well-drained sands, either on outwash plains or on shallow lake bottoms. The hardwood lands include the Valders till exposures and sandy soils with till near the surface. The conifer-hardwood area consists of poorly drained sandy mineral soils while the swamp includes the organic soil deposits.

A detailed and high-quality soil survey was published for Cheboygan County by Foster et al. in 1939. The soil units mapped provide a generally satisfactory key to the distribution of till, outwash sands, glacial deposits, and other geological materials. Unfortunately, however, soil series names are anything but definite or static. A revised and improved classification of Michigan soil series was printed in 1952 by E. P. Whiteside and I. F. Schneider of Michigan State College. They use essentially a two-way classification based on texture of the parent material and soil order (which represents the effect of drainage on profile development).

The present discussion is an effort to reconciliate the older classification scheme used in Cheboygan County with recent changes, and to show the relationship between the soils as mapped and geology. The mineral soils of the two counties can be classified on a geologic basis as having been derived from (1) Valders glacial till, (2) Valders outwash, (3) lake deposits, and (4) outwash overlying till or lake deposits. Organic soil types are not considered.

Valders till soils.—The basic Valders till soil developed under well drained conditions is the *Onaway* series. This occurs predominantly as a loam. Inasmuch as the Valders ice moved in from across the Lake Michigan basin, it carried substantial quantities of lake sand and in places is more of a sandy loam or even a loamy sand. These are mapped as *Emmet sandy loam*. This type, however, is also mapped where outwash sand overlies Valders loam. Moreover, *Emmet loamy sand* is mapped for the better outwash soils. This latter type has since been renamed *Wexford*. Even so, *Emmet* as currently mapped is still a hybrid soil referring both to a sandy till and to sand over till.

Poorly drained conditions are not commonly found above the Lake Algonquin shores. Below this, they commonly result from planation of Valders till by lake waters. In Cheboygan County, imperfectly drained Valders till deposits below Algonquin levels are mapped *Selkirk silt loam*. Poorly drained soils with humic-gley horizons are mapped as *Bergland clay loam* or *Bruce very fine sandy loam* whether they are developed on Valders till or lake deposits. According to the present classification, the Valders till portion of these soils would currently be mapped as *Mackinac* under imperfect drainage and *Angelica* under poor drainage.

Lake soils in the area are more commonly developed from a two-storied parent material than from a single fine-textured one. The clays and silts deposited under the higher lake levels or under frozen surfaces are commonly overlain with later sands. Under these conditions, *Iosco sandy loam* is mapped under imperfectly drained conditions, *Ogemaw sandy loam* under wetter ground-water podzol conditions, and *Munuscong sandy loam* under poorly drained humic-gley conditions.

Most of the lake deposits in Cheboygan and Emmet counties, however, are shallow-water sands. The well-drained upland sands—and most fall in this class—are mapped as *Grayling*, *Rubicon*, or *Kalkaska* in increasing order of percent of fine particles and profile development. *Saugatuck*, *Granby*, and *Newtown* are mapped for the imperfectly drained and poor conditions as with the outwash sands.

Finally, recent sand deposits and dunes are classified as *Eastport*, *Weare*, or *Wallace*, again depending on the amount of soil profile development. *Eastport* is the azonal soil and *Wallace* the podzol.

The above classification and reconciliation is summarized in table 2. It accounts for over 95 percent of all the mineral soils mapped in Cheboygan County.

TABLE 2

Soil Catenas of Cheboygan County

(Values in parentheses refer percent of county mineral soil area in soil type)

Parent material	Soil drainage		
	Well drained	Imperfectly	Poorly drained
Till	Emmet sandy loam (in part) & Onaway loam (4.1)	Selkirk (1.4) (now Mackinac)	Bergland and Bruce (in part, now Angelica)
Non-stratified outwash	Roselawn (7.0) & Emmet loamy sand (now Wexford, 10.7)	Saugatuck (in part)	Granby & Newtown (in part, now Roscommon)
Outwash over layer of till	Emmet sandy loam (in part, 19.6)		
Lake silts	Bohemian	Brimley	Bergland & Bruce (in part, 2.2 & 0.5)
Lake sands over silts and clays	Emmet (in part, now Menominee)	Iosco (0.6) & Ogemaw (5.5)	Munuscong (7.0)
Lake sands	Grayling (2.4), Rubicon (11.5), & Kalkaska (9.6)	Saugatuck (in part, 4.1)	Granby & Newtown (in part, now Roscommon, 5.9)
Recent sands and dunes	Eastport (1.2) Weare (0.1), & Wallace (0.5)		

Outwash soils.—The proglacial outwash formed non-stratified or poorly stratified deposits in which sands predominate but in which gravels and some silt and clay are also found. Where the content of fines is low, podzolization is minimal and the soils are mapped as belonging to the *Roselawn* series. Where enough fines were present to permit distinct podzolization under upland conditions, *Emmet loamy sand* (now *Wexford loamy sand*) was mapped in Cheboygan County. As with the till deposits, poor drainage conditions are largely found where these outwash deposits have been planed by subsequent lake stages. In this case again, no distinction is made as to whether the resulting sands were of outwash or of lake bottom origin. *Saugatuck sand* is mapped under imperfect drainage conditions, and both *Newtown sand* (acid) and *Granby sand* (limy) under humic-gley conditions. The latter two series would be grouped as *Roscommon sand* under the present classification.

Lake deposits.—Lake deposits vary from compact clays deposited in perfectly still waters (i.e. at deep levels or under ice) to coarse sands deposited in turbulent waters (near the shore). Due to their topographic location, they are usually imperfectly or poorly drained, and the well-drained members of the soil catena are apt to be either not present or of too little importance to be mapped. *Bergland clay loam* is the heaviest lake deposit mapped in Cheboygan County followed by *Bohemian very fine sandy loam* (well drained), *Brimley very fine sandy loam* (imperfectly drained), and *Bruce very fine sandy loam* (poorly drained or humic-gley). *Selkirk silt loam* as mapped in Cheboygan County may well include lake deposits.

RECENT GEOLOGIC HISTORY

The geologic history of the Cheboygan-Emmet area from Valders time to the present is summarized in table 1. In so far as the evidence in the two counties provides a check, the late Wisconsin chronology agrees perfectly with that described for the Lake Michigan basin on the basis of studies near South Haven (Zumberge and Potzger, 1955). The Valders glaciation maximum may be dated by the Carbon 14 determinations of Two-Creeks wood at somewhat less than 11,000 years ago. At its major recessional position filling the upper Michigan basin, meltwater was carried across the Emmet-Cheboygan area into the Huron basin. This could well have been contemporaneous with the Calumet stage of Lake Chicago. The pre-Algonquin shores here described could well represent the *Toleston*[1] level while the major Algonquin shore is, of course, that of the highest and most clearly defined level of Lake Algonquin. In this case, it is probable that the Toleston stage of Lake Chicago did not develop at the 605-ft. level south of the hinge line, but rather that it developed at a lower level in that area. Otherwise, it would be necessary to postulate down-warping of the Toleston beach, an improbable occurrence. This lake existed until about 8000 years ago in a wet and probably a cool climate as it is correlated with the presence of a spruce-fir forest. Following the lowering of Lake Algonquin due to the opening of the North Bay outlet, the waters in the Lake Michigan basin stood at the Chippewa level, the principal manifestation in Cheboygan and Emmet counties being the steeply graded stream valleys of Maple River, Nigger Creek, and the Little Traverse river system. The Nipissing stage, dated at about 3500 years by Hough (1953) and at a little less than 4000 by Zumberge and Potzger occurred during a dry climatic era and probably coincides with the Xerothermic time of the pollen analysts. The general pollen profile from Douglas Lake (Wilson and Potzger, 1943) agrees with this general sequence.

[1] Spelling now Tolleston. We retain Leverett's spelling.

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SOME ROCKS OF THE CENTRAL ST. ELIAS MOUNTAINS, YUKON TERRITORY, CANADA

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ABSTRACT. Brief observations were made on scattered bedrock outcrops in the heart of the St. Elias Mountains incidental to glaciological work on upper Seward Glacier. The principal rocks in this geologically unexplored area are metamorphic and representative of the amphibolite facies of medium-grade regional metamorphism. Four units are distinguished as follows: Unit I is chiefly a coarse biotite or hornblende gneiss with interlayers of micaceous quartzite and coarse mica schist. Unit II comprises a variety of schists, principally quartz-biotite schist, hornblende-biotite schist, and hornblende schist with some micaceous quartzite, amphibolite, lime-silicate rock, and at least one bed of metaconglomerate. Unit III is predominantly dark amphibolite and light foliated marble with minor amounts of chloritic hornblende schist, hornblende-biotite schist, quartz-biotite schist, and lime-silicate rock. Unit IV is a uniform medium- to coarse-grained hornblende gneiss. Units I, II and III are predominantly, if not wholly, of metasedimentary origin and comprise in each instance beds totaling at least 1000 to 2000 feet in thickness and possibly much more. Unit IV is possibly a pre-metamorphic diorite and quartz-diorite intrusive. These rocks are all definitely older than Early Cretaceous, with a Paleozoic age slightly favored by indirect evidence, although either an early Mesozoic or Precambrian age is almost as likely.

At the east end of the Mt. Logan massif are some dark calcareous graywacke, sandy limestone, calcareous quartzite, and conglomerate beds which contain earliest Lower Cretaceous fossils. These rocks are younger than the metamorphics.

The above rocks, especially the metamorphics, are intruded by small bodies of quartz diorite and by still younger lamprophyre dikes. These intrusives are younger than Early Cretaceous and by inference older than Paleocene.

Northwesterly structural trends in the metamorphics are largely overshadowed in their influence on the present topography by larger and younger westward-trending structures which by analogy with better-known relations on the south face of the St. Elias Mountains probably involve Laramide(?) folding and later movements along steep thrust faults from late Pliocene to the present.

INTRODUCTION

During the summers of 1948 and 1949 glaciological studies on upper Seward Glacier afforded opportunity for brief observations of scattered bedrock exposures in the heart of the St. Elias Mountains along the Alaska-Yukon border. This remote, little-known area is a blank on the latest geological maps of North America and Canada. For this reason and because possibilities of further geological work in the area appear remote, the few data gathered are placed on record.

The area treated is along the eastern and northeastern margin of a large ice-filled intermontane valley in the Ice Field Ranges of the St. Elias Mountains (Bostock, 1948, p. 98-100), Yukon Territory, Canada (N. Lat. 60° 25′ W. Long. 140° 00′). The surface of upper Seward Glacier is 5500 to 7000 feet in altitude, and the adjacent peaks include Mt. Logan (19,850) to the northwest, Mt. St. Elias (18,008) to the west, and Mt. Vancouver (15,700) to the east. Geophysical work indicates that the ice in the upper Seward intermontane valley is 2000 to 4000 feet thick, so total bedrock relief is roughly 16,000 to 17,000 feet.

Bedrock is exposed on precipitous slopes bordering this valley and in small rock islands or nunataks projecting through the cover of ice and snow (pl. 1-A). Observations were made at localities scattered over an area about 17 miles north-south by 11 miles east-west (fig. 1). This area was not mapped

in the usual sense, for lateral tracing of contacts and "walked out" correlations were not possible owing to the extensive cover of ice and snow, the inaccessible and isolated nature of many exposures, the steepness and instability of the

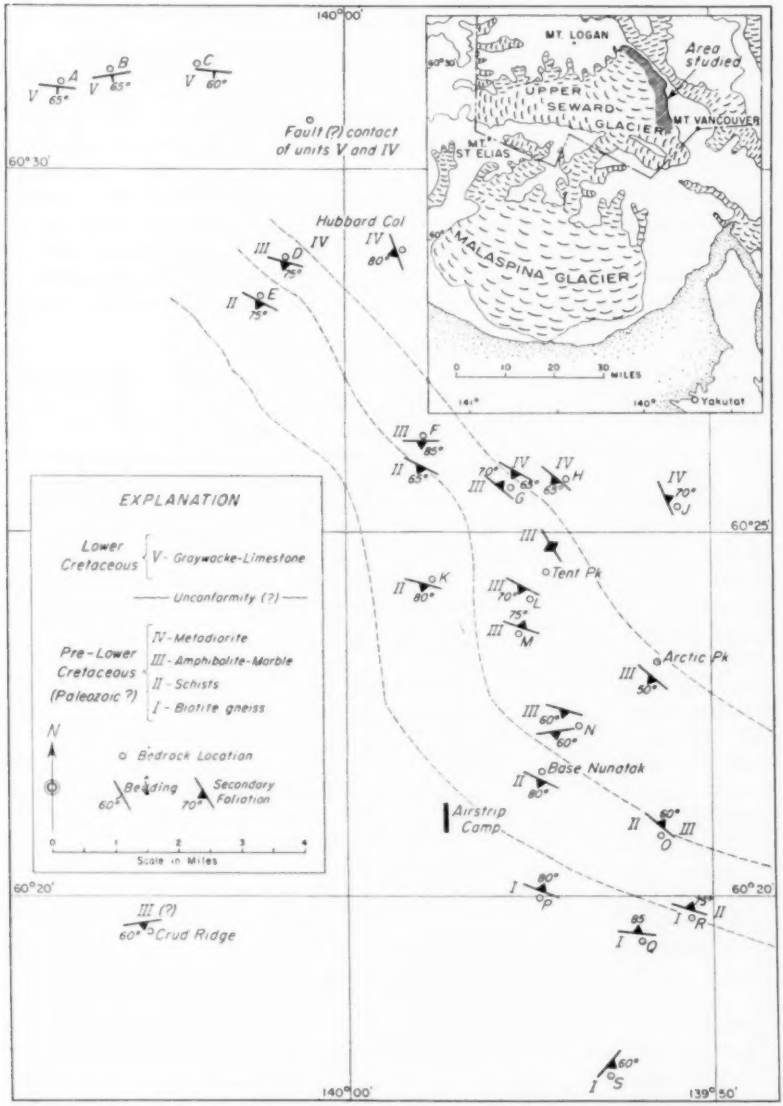


Fig. 1. Bedrock outcrop map along eastern and northeastern margin of upper Seward Glacier.

slopes, and limitations of time. However, reasonably complete rock collections were made, and many specimens have been studied in thin section.

The map (fig. 1) is based upon triangulation data provided by Walter A. Wood supplemented by a perspective grid constructed from a U. S. Air Force oblique trimetrigon air photo. The major control is good, but the order of accuracy on details is low. Boundaries sketched between metamorphic units are diagrammatic.

Opportunity to undertake this work and logistical support were provided by "Project Snow Cornice," an endeavor of the Arctic Institute of North America directed by Walter A. Wood. We are greatly indebted to Mr. Wood and all members of the project in 1948 and 1949, especially F. B. Leighton. The Engels, Celeste G. and Albert E., have aided in laboratory study and interpretation of the rock collections. Don J. Miller and George Plafker offered constructive comments on the manuscript. Noel E. Odell (1951) independently investigated some of the bedrock in remote and higher parts of this region.

METAMORPHIC ROCKS

General statement.—In area of exposure and lithologic variety metamorphics are by far the principal rocks in the region, and they are the oldest. Observations during airplane flights suggest that similar rocks underlie much of the surrounding country for tens of miles to the north, east, and west. Along the south face of the St. Elias Mountains they are brought into contact with Mesozoic and Tertiary sedimentary rocks composing the lower slopes of the range by a large east-west thrust fault (Russell, 1891, p. 168, 174; Tarr and Butler, 1909, p. 149; Miller, 1951, p. 28; Plafker and Miller, 1954, p. 12). Russell's (1891, p. 167, 173-174) St. Elias schist is undoubtedly part of this metamorphic sequence as it is lithologically similar and geographically adjacent (Filippi, 1900, p. 235; Plafker and Miller, 1954, p. 4), but we see little reason for applying this name to the entire group. The upper Seward rocks are representative of the amphibolite facies of medium-grade regional metamorphism. Four units are distinguished on the basis of lithology, all are metasedimentary except Unit IV which is possibly igneous. The sequence of units is consistent throughout the area studied, but so little is known about structure that a top or bottom cannot be determined with confidence. These rocks are intruded by small bodies of quartz diorite and by lamprophyre dikes which are later than the principal episode of metamorphism.

Local field evidence definitely shows that the metamorphic rocks are older than Early Cretaceous, but there is little basis for choosing among an earlier Mesozoic, Paleozoic, or Precambrian age. Rocks of similar lithology and corresponding degree of metamorphism in neighboring parts of Canada are consistently assigned to the Yukon group which is generally regarded as Precambrian but may include some Paleozoic rocks (Cairnes, 1915, p. 69-71; Bostock, 1952, p. 15-24; Kindle, 1952, p. 27-29; Muller, 1954, p. 1-2). Most Paleozoic rocks within a radius of 100 miles of the upper Seward region are less metamorphosed (Cairnes, 1915, p. 71; Capps, 1916, p. 31-47; Moffit, 1938, p. 22-29; Bostock, 1952, p. 17-22; Kindle, 1952, p. 29-31; Sharp, 1943, p. 630-633; Muller, 1954, p. 2), the few exceptions being instances in which

assignment of a Paleozoic age is not strongly supported by direct evidence (Mendenhall, 1905, p. 30; Tarr and Butler, 1909, p. 147; Moffit, 1938, p. 25). However, it has been shown repeatedly that degree of regional metamorphism is such a poor basis for age determination that a possible Paleozoic age for the upper Seward rocks cannot be eliminated on this score.

Plafker and Miller (1954, p. 3) suggest that metamorphics on the south face of the St. Elias Range, certainly a part of the sequence described herein, are Mesozoic and older(?), and Kennedy and Walton (1946, p. 68-69) favor a Mesozoic age for similar rocks in the Lituya Bay-Mount Crillon area, 150 miles southeast. In neither instance is the evidence direct nor compelling, but the southeastward trend of geosynclinal belts in this region (Payne, 1955) is a point in favor of such a correlation. However, the abundance of Paleozoic sedimentary rocks in neighboring areas and the lack of any marked regional metamorphism in nearby Permian sedimentary and volcanic rocks (Moffit, 1938, p. 31, 35; Sharp, 1943, p. 633-634; Muller, 1954, p. 2) suggest to us that the metamorphics of the upper Seward area are possibly Paleozoic.

Unit I—Biotite gneiss.—This unit was seen only in the southeastern part of the area studied, specifically at locations P, Q, R, and S (fig. 1). At location R it passes with sharp but seemingly conformable contact into Unit II. Unit I is arbitrarily placed at the bottom of the metamorphic sequence (fig. 2), but in so far as direct evidence goes it may actually be at the top.

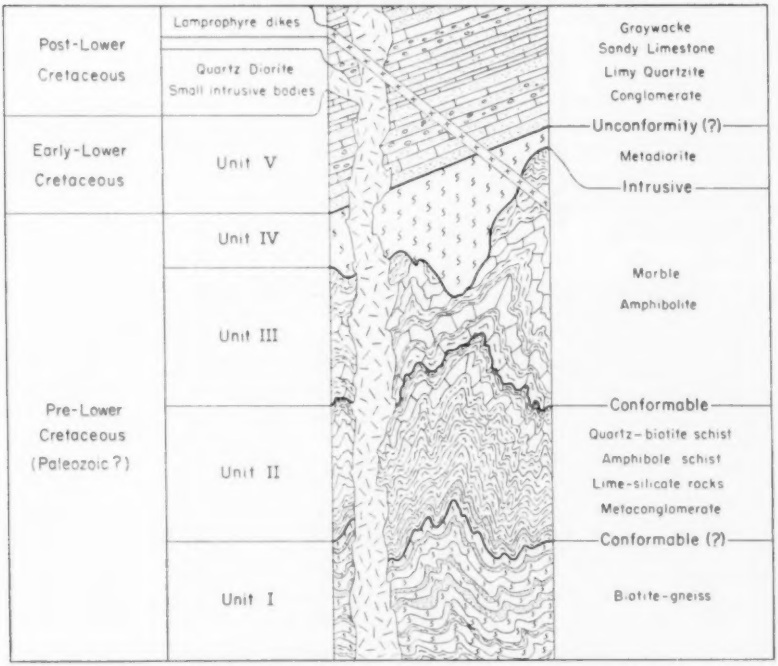


Fig. 2. Columnar section of rock units exposed in map area. (fig. 1).

If the rocks of this unit are not repeated by folding, then they are at least 5000 feet thick. However, close folding is suggested by the conformity between foliation and compositional variations, probably related to original layering, so the thickness is conservatively estimated at 1000 to 2000 feet.

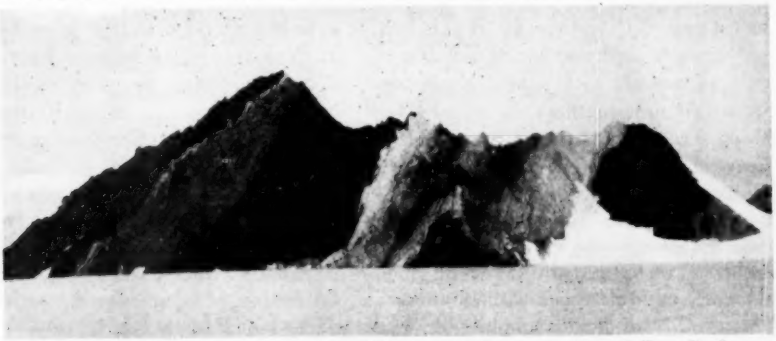
Lithologically, this is the most uniform and homogeneous unit within the metamorphic series. It is principally a coarse biotite or hornblende gneiss with thin interlayers of micaceous quartzite and coarse mica schist. Locally the gneiss and schist are garnetiferous, and in places both the biotite and garnet form nodular aggregates. A representative sample of gneiss near the middle of the unit has the following estimated percentage mineral composition: 40 quartz, 20-30 oligoclase, 15 biotite, 5-10 hornblende, and some apatite, sphene, zoisite, chlorite, and zircon. Modal analysis (300 counts) of a specimen near the top of Unit I (loc. R) gives: 43.5 quartz, 41.5 andesine, 13 biotite, 0.8 sphene, 0.6 magnetite, and 0.6 muscovite or sericite. Another rock nearer the contact with Unit II (loc. R) has by modal analysis (300 counts): 31 percent quartz, 26 oligoclase, 19 green hornblende, 10 zoisite, 6 diopside, 5 muscovite, 1 magnetite, 0.5 apatite, and 1.5 unidentified. The mineralogy and compositional layering observed in the rocks of this unit suggest that they were derived from sedimentary materials such as graywacke or tuff or from slightly calcareous sandy ferruginous shale and impure argillaceous sandstone with introduction of Na_2O or removal of K_2O.

Unit II—Schists.—This unit comprises a narrow belt trending north-northwestward most of the way across the area studied (fig. 1). It is best exposed at locations E, F, K, O, R, and the Base Nunatak, the last being most representative. The trend of the belt is slightly oblique to the predominant attitude of the foliation which bears a bit more westerly and dips 75 to 80 degrees southwesterly. This may be due to the superimposition of small folds and contortions, such as have been observed in outcrops, upon larger folds. Contact with Unit I is sharp, but with Unit III it is gradational and is expressed chiefly by an increase in calcareous material. The variety and thickness of individual rock types observed suggest that this unit cannot be less than 1000 feet thick, and it may actually be several thousand feet.

Unit II is composed primarily of schists and finely banded gneisses which for the sake of distinction from other gneisses in the area are all termed schists, the most abundant being: quartz-biotite schist, hornblende-biotite schist, and hornblende schist. Also present are micaceous quartzite, amphibolite, lime-silicate layers containing diopside and zoisite, dark to light gray foliated marble of sugary texture, and one or more beds of metaconglomerate. Lenses and layers of marble, lime-silicate rock, and amphibolite become more abundant toward the contact with Unit III. Some of the schists are garnetiferous, and porphyroblasts of andalusite were seen in one layer. Retrograde metamorphism has locally produced chlorite and chlorite-sericite schists. Metaconglomerate is suspected at several localities, but alteration is so great that recognition is uncertain except at location K where 50 feet of unquestionable metaconglomerate are exposed. It contains stones up to 6 inches in diameter of gneiss, a fine granitic rock, black fine schist, black coarse am-

PLATE 1

A. Air view northwestward along rock ridge bounding northeast edge of upper Seward Glacier. Localities lettered to correspond with map (fig. 1). Photo by W. A. Wood.



B. Amphibolite and marble of Unit III as exposed on east face of Tent Peak.

C. View west-northwestward into northwest corner of map area (fig. 1). Metadiorite of Unit IV on right and early Lower Cretaceous sedimentary beds of Unit V compose dark ridge in center.

phibolite, quartz, and a gabbroic rock all set in schistose matrix. The individual stones are elongated and have modified structure and texture.

Modal analyses (300-500 counts) were made of three specimens from Unit II with the following results: a biotite-hornblende schist from the Base Nunatak has 23 percent quartz, 32 andesine, 19 plagioclase or quartz (probably mostly the latter), 18 biotite, and 8 hornblende; a hornblende-biotite schist from location O has 27 quartz, 43 andesine, 8 biotite, 20 hornblende, and 2 accessory; a badly altered lime-silicate rock near the base of Unit II (loc. R) has 26 quartz, 38 plagioclase highly altered to sericite and calcite, 4 potash feldspar, 23 diopside, 4 hornblende, 2 sphene, 2 zoisite, 1 apatite, and a trace of opaque minerals.

The composition and interlayering of these rocks show that this unit is metasedimentary and probably consisted originally of conglomerate, sandstone, argillaceous sandstone, sandy shale, dirty carbonate layers partly argillaceous, tuffaceous, and sandy, and of relatively pure limestone.

Unit III—Amphibolite—marble.—These rocks compose a belt extending north-northwestward across the study area on the northeast side of the schist belt (Unit II). Narrowing of this belt in a northwesterly direction is probably due to the crosscutting relation of Unit IV. The contrast between light-colored marble and nearly black amphibolite makes this the most easily identified unit in the area (pl. 1-B). It is prominently displayed on Arctic Peak, Tent Peak, and along part of the bedrock ridge separating the upper Seward and Hubbard glaciers. Most of the rocks at locations D, F, G, L, M, N and O belong to this unit, and those composing Crud Ridge in the southwestern corner of the study area are tentatively included in it, although they may actually belong to a wholly different but lithologically similar unit.

The rocks of Unit III show intense deformation and much flowage of the carbonate materials with thinning and thickening of layers. During such flow, bodies of brittle amphibolite were detached and carried along in the marble. Small-scale tight folding is locally visible, and larger folds are inferred from the distribution of marble and amphibolite layers. The contact with Unit II looks to be of gradational sedimentary nature, but Unit IV may be intrusive into Unit III. Deformation makes an estimate of thickness for Unit III little more than a guess. In spite of considerable folding, the individual exposures indicate a thickness of at least 1000 feet, and the relatively great width of the belt underlain by this unit suggests a possible total thickness of several thousand feet.

Unit III consists predominantly of amphibolite and foliated marble with small amounts of chloritic hornblende schist, hornblende-biotite schist, quartz-biotite schist, and lime-silicate rocks. The amphibolites are black to greenish, mostly fine-grained, and locally mylonitic to schistose. Some have radial aggregates of clinozoisite, a few contain actinolite, and most include some calcite plus a little epidote, quartz, and sphene. Retrogression to chlorite has occurred. A content of 40 to 50 percent hornblende is not unusual, and as much as 80 percent hornblende was observed in thin sections of some amphibolites.

An amphibolite from location N, by modal analysis (300 counts), contains 25 percent quartz and andesine, 71 hornblende, 3 magnetite, and 0.7 biotite; and another amphibolite from the same locality contains 19.5 percent quartz and plagioclase, 72 hornblende, 6 nodules (unidentified), and 2.5 magnetite.

The marble layers are mostly light colored, and though clearly recrystallized, they are not especially coarse-grained. Some are mottled gray and white, but a streaking of these colors parallel to the foliation is more common. Variations range from relatively pure calcite marble through marble containing some diopside, to lime-silicate rocks consisting of calcite, quartz, pyroxene, amphibole, and epidote. Grossularite, epidote, and quartz have been developed along contacts with intrusive quartz diorite. Talc was observed in marble near the contact with Unit IV at the north end of location D and also in nearby lime-silicate rocks where it had developed as an alteration of tremolite. Lime-silicate rocks are abundant on Crud Ridge, and one sample studied contained large crystals of tremolite, possibly some actinolite, chlorite, serpentine, magnetite, secondary quartz and pyrite, and veinlets of celestite.

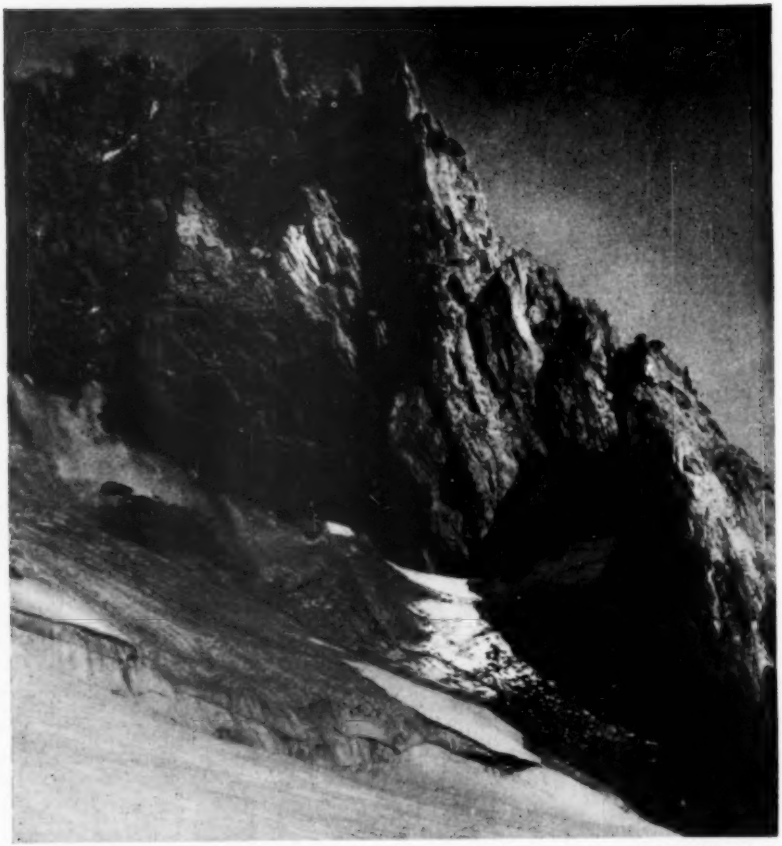
Unit III has been derived principally from carbonate-rich sedimentary rocks ranging from relatively pure calcite limestone to impure shaly limestone or calcareous shale. No direct evidence that the section contained tuffs or lavas was found, although similar amphibolites from nearby areas are possibly of volcanic origin (Plafker and Miller, 1954, p. 3; Kennedy and Walton, 1946, p. 68).

Unit IV—Igneous gneiss.—This unit crops out on nunataks and ridges in the northeastern part of the study area. It was seen at Hubbard Col, the north ends of locations D and G, and is a principal constituent at locations H and J. Debris on a glacier draining the west face of Mt. Vancouver was derived almost wholly from this unit, and specimens collected by M. M. Miller along the northwest ridge of Mt. Vancouver suggest that much of this peak consists of Unit IV. The higher part of Mt. Logan may be composed of similar rock (Lambart, 1926, p. 600).

The contact between Unit IV and Unit III is thought to be intrusive, although the field relations are not completely clear. Close observation of the contact has been possible over short distances in only two places. On the map (fig. 1) Unit IV appears to cut across Unit III in a northwesterly direction, so that it lies in contact with different layers within that unit. In local exposures, the contact zone is irregular (pl. 2), and crosscutting relations with apophyses of gneiss extending into Unit III have been observed. Tremolite and talc have been developed in the carbonate rocks of Unit III near the contact. The gneiss of Unit IV is well foliated, and since the foliation is essentially conformable with that of the other metamorphic units, it was presumably emplaced prior to metamorphism.

The principal rocks of Unit IV are a uniform medium- to coarse-grained hornblende gneiss and a quartz-hornblende gneiss that can easily be metamorphosed diorite and quartz diorite. Modal analysis (500 counts) of one specimen from Hubbard Col gives 39 percent quartz, 46 plagioclase, probably oligoclase, 10 biotite, 3 hornblende, 0.5 apatite, 0.5 magnetite, and traces of

PLATE 2



Pre-metamorphic intrusive(?) contact between dark metadiorite of Unit IV (left) and light marble-lime silicate rocks of Unit III (right) at north end of location D (fig. 1).

zircon and zoisite. Small to moderate-sized bodies of amphibolite and metasedimentary gneiss in Unit IV probably represent inclusions in the original rock. The gneiss also contains small dikes and pods of alaskite, in places pegmatitic, which look to be pre-metamorphic and are probably related to the original dioritic intrusive. Epidote veinlets are also abundant.

Unit V—Cretaceous sedimentary rocks.—The western half of a low ridge extending east from the base of Mt. Logan to Hubbard Col is composed of dark-colored, well-bedded sedimentary rocks (pl. 1-C). Air observations and photographs suggest that these same rocks extend west along the south base of the Mt. Logan massif. Their structure, where studied, is a simple homocline trending roughly westward and dipping 60° to 65° south. Local deviations are

probably related to faulting. The contact on the east with the metamorphic rocks is well exposed, appears to be nearly vertical, and is probably a fault. Unfortunately, it is not easily accessible and has not been studied at close hand.

The rocks are chiefly black and brown calcareous graywacke, sandy limestone, calcareous quartzite, and conglomerate in that order of abundance. The limestone and graywacke are interbedded, and at least some and perhaps all of the calcite in them is of secondary detrital origin. These rocks are cut by veinlets and pods of coarsely crystalline white and black calcite and are cemented by secondary calcite. The quartzites are slightly micaceous and range in grain size from very fine almost to grits. The fine-grained quartzite is jet black, and the coarser quartzite is brownish and gray. The conglomerate layers contained well-rounded stones up to 9 inches in diameter of granitic rocks, black limestone, gneiss, black chert, and black quartzite.

These rocks are only slightly metamorphosed. Biotite in some of the quartzite and graywacke layers may be secondary. Quartz veinlets, quartz-pyrite nodules, and dikes of quartz diorite cut the rocks. The thickness in the area studied is at least 1000 feet, and the total thickness may be several times that judging from the attitude and distribution of the beds.

Fossil pelecypods (aucellas) were collected in limy graywacke and limy quartzite beds in two parts of location B, at about the same stratigraphic position. These fossils were studied by Imlay (Imlay and Reeside, 1954, p. 229) and comparisons made with *Aucella okensis* Pavlow, *A. subokensis* Pavlow, and *A. volgensis* Lahusen. "If correctly identified, these would represent the Berriasian stage of the very base of the Lower Cretaceous." Regardless of specific identity, these aucellas characterize the earliest Cretaceous of the Boreal region and the Pacific coast of North America.

It seems most probable that these rocks are correlative to some part of the 12,000 feet of Lower Cretaceous beds in the Dezadeash group, 65 miles to the east (Kindle, 1952, p. 35-37), which are lithologically similar and contain corresponding fossils. Lower Cretaceous sedimentary rocks are also widely distributed in the Chitina Valley, 125 miles to the northwest, and, although not identical lithologically, they contain the same fossils (Moffit, 1938, p. 70-89). The Yakutat group (Plafker and Miller, 1954, p. 5-6) of Cretaceous age, along the south face of the St. Elias Range, may include related rocks.

IGNEOUS ROCKS

The metamorphic rocks of possible intrusive origin composing Unit IV have already been described. The rocks treated here were intruded after the principal episode of metamorphism and are, in all probability, post-Early Cretaceous. The intrusive bodies consist of quartz diorite and lamprophyre and are similar to intrusives observed in the crystalline rocks closer to the coast (Plafker and Miller, 1954, p. 3-5).

Quartz diorite.—This is the most abundant intrusive rock in the study area, although it consitutes not more than 5 to 10 percent of the total rock exposure studied. Representative examples were seen on the base Nunatak and at location N. This rock forms dikes a few to 50 feet wide and larger irregular

discordant bodies of undetermined size and shape. The quartz diorite intrudes mostly Units I, II, and III, but quartz-diorite dikes also cut the Lower Cretaceous beds of Unit V. Some exposures of quartz diorite have a strong system of closely spaced joints, and in places the rock displays a distinct planar structure owing to oriented biotite or amphibole crystals, but this is usually near contacts. Oriented hornblende laths locally give a prominent lineation.

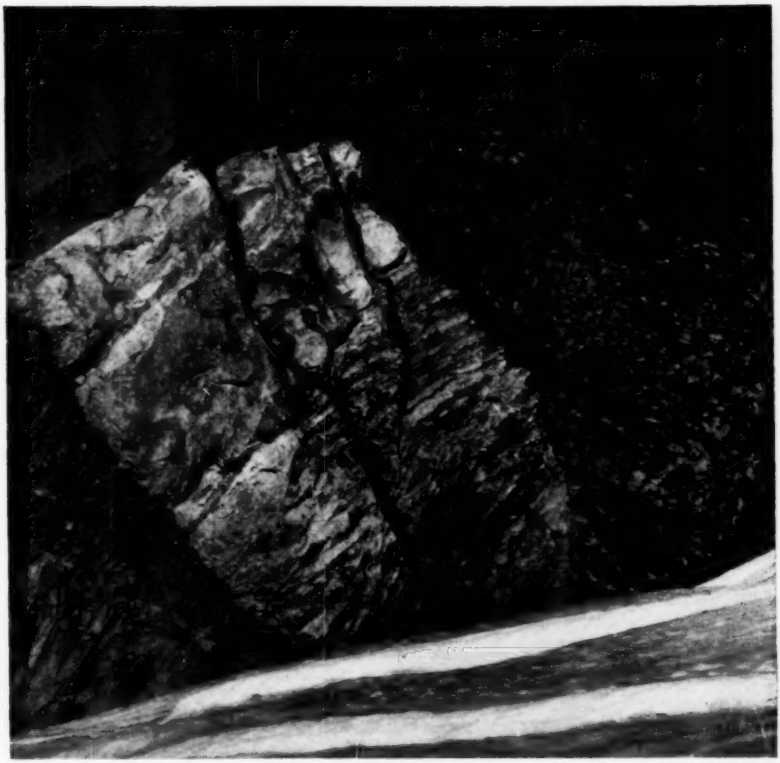
The larger bodies are composed of uniform, medium- to fine-grained equigranular quartz diorite of light gray to greenish-gray color. The varietal minerals differ from place to place and are biotite, biotite and hornblende, hornblende, and hornblende and muscovite in about that order of abundance. The muscovite may be an alteration product, as the rock specimen in which it appears also contains considerable secondary calcite. Potash feldspar was identified in only two thin sections where it composed not more than 0.5 percent of the rock. Some of the smaller dikes are porphyritic, and the plagioclase feldspar phenocrysts are zoned. Modal analysis (1000 counts) of a specimen from the Base Nunatak gives 30 percent quartz, 52 andesine, 0.5 potash feldspar, 14 biotite, 2.5 hornblende, 0.5 muscovite, 0.5 zoisite and traces of apatite, zircon, and sphene; and a specimen from location N contains 30 quartz, 59 andesine, 8 biotite, 2 chlorite, 1 muscovite, and a trace of potash feldspar.

The larger bodies contain inclusions of the metamorphic rocks in various stages of digestion. Chilled borders were observed, along intrusive contacts. In addition to quartz and quartz-chlorite veins, pods and dikelets of leucophyre and of pegmatite are locally abundant within the quartz diorite and in the wall rock. Some pegmatite contains small books of biotite and muscovite, and locally black tourmaline.

Lamprophyre.—Hornblende lamprophyre dikes, a few inches to a few feet across, are widely distributed in the study area. They intrude the metamorphic rocks and the quartz diorite, and the age relations are clearly established in some exposures (pl. 3). Texturally the lamprophyre ranges from fine to locally coarse, and some of the larger, coarser-grained dikes have chilled borders. Hornblende and plagioclase are the principal minerals with magnetite, sphene, epidote, clinozoisite, and chlorite among the accessory or secondary minerals. Up to 40 percent hornblende, both brown and green, is common, and one dike contained 80 percent hornblende as large crystals. A specimen of lamprophyre from the Base Nunatak has 20 percent plagioclase, 50 hornblende, 13 magnetite, 11 chlorite, 5 sphene, and 1 apatite; and another from location N shows 47 hornblende, 41 groundmass of feldspar, chlorite and hornblende(?), 7 chlorite, and 5 magnetite. Close geographic association suggests that the lamprophyres are genetically related to the quartz-diorite intrusives.

Age of igneous rocks.—The igneous rocks described are clearly younger than the metamorphics, and their intrusion occurred subsequent to the principal episode of metamorphism. Quartz-diorite dikes, similar in all respects to other quartz-diorite bodies in this area and presumably related thereto, intrude the early Lower Cretaceous beds, so both the quartz diorite and the lamprophyre are considered to be younger than Early Cretaceous. The lack

PLATE 3



Block of marble from Unit III included in quartz-diorite intrusive and cut by dark lamprophyre dikes, location N.

of intrusives in Tertiary rocks, possibly as old as Paleocene, at the south base of the St. Elias Mountains (Miller, 1951, p. 6, 11) suggests that the quartz diorite and lamprophyre are pre-Tertiary. Their intrusion probably occurred sometime within the Cretaceous or in the Cenozoic-Mesozoic interval. However, nothing definite is known concerning possible relation to the Coast Range batholith or to somewhat younger bodies (Laramide) farther west and north (Moffit, 1933, p. 158; 1938, p. 106-107; Moffit and Pogue, 1915, p. 58-59; Capps, 1933, p. 286).

STRUCTURAL TRENDS

Two principal structural trends can be seen or inferred within the area studied. The first is a northwesterly grain displayed by foliation within the metamorphics and expressed in the orientation of individual ridges and canyons controlled by this structure. This structural trend was established before Early Cretaceous time and may represent pre-Permian Paleozoic deformation.

The second and younger structural grain is inferred from major westward-trending features such as the great trough in which the upper Seward Glacier lies, the high Mt. St. Elias-Mt. Cook ridge to the south, and the elongate Mt. Logan massif to the north. This structural trend extends as a prominent feature far westward into the Chugach Mountains but gives way eastward to southeasterly trending structures. In the area studied nothing is known firsthand of its nature. By analogy with better-known features of similar trend on the south face of the St. Elias Mountains, it seems likely that movements on steep thrust faults are involved beginning in late Pliocene to early Pleistocene time and continuing to the present (Miller, 1951, p. 28; Plafker and Miller, 1954, p. 12-13). A west-trending grain may also have been initiated by an earlier episode of folding (possibly Laramide) that can be inferred from evidence in the coastal belt.

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THE UNUSUAL BENDING OF SOME MARBLE HEADSTONES

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ABSTRACT. Certain thin-tabular marble headstones in the Grove Street Burial Ground, New Haven, Connecticut, are curved several degrees from their former vertical position, and some of the headstones are warped to a saucer-like form. There is no apparent relation between the age of the stones and either degree or commonness of curvature. The writers believe partial disaggregation of interlocking calcite grains the most likely of several possible explanations for the curvature.

INTRODUCTION

The pronounced curvature of vertical, thin-tabular marble headstones in the Grove Street Burial Ground, New Haven, Conn., led the writers to conduct a brief investigation of the nature of the bending and its origin. The sagging of horizontal marble stones under the influence of gravity is well known, but this is the first example known to the writers of bending in vertical headstones.

Although the stones examined may not all have had the same history, the Superintendent of the Burial Ground, Mr. Malcolm Munson, believes that most of the marble came from Italy as ballast for sailing vessels during the early and middle 19th century.

INVESTIGATION AND RESULTS

The stones studied were all of white crystalline marble similar to that in the curved stones, and of shape and size considered optimum for bending: thin tablets not less than 3 feet high nor more than 0.2 feet thick. Few stones shorter than 3 feet are curved, but some tall stones more than 0.2 foot thick do show curvatures of several degrees. The Burial Ground is divided into eight large blocks; in two of these all the stones meeting these limitations were studied. A spot check was made of similar stones having unusual features in the remaining six blocks of the cemetery. Since no material could be obtained for microscopic examination, the writers' conclusions are based entirely on megascopic observations.

The white marble of the headstones is equigranular but ranges in grain size in different stones from 0.2 to 2 mm. The most common grain size is about 1 mm. The rock is mainly pure marble, but some stones contain several percent of amphibole (tremolite ?), quartz, and mica. These accessories commonly are arranged in sheets that emphasize the foliation of the marble, which is common in some stones. The stones have been cut with little regard for any linear or planar properties, and the sporadic foliation seems not to influence the curvature. Cross-fractures are found only in curved stones and are approximately radial to the curvature.

Most of the headstones are simply bent, but a few are warped into shallow, saucer-like forms. The curvature ranges to about 12 degrees, but most commonly is only a few degrees. Probably most stones collapse when curved beyond about 8 degrees. Generally only the lower half of the stone is curved.

Of 122 stones examined, 45 are bent more than one degree, and 10 are bent more than 5 degrees. The depression in the most strongly warped stone

PLATE 1

A B

A. Marble headstone dated 1858 with a 10-degree bend. The nearly radial fracture does not extend far from the edge of the stone.

B. Bent tombstone (fourth in line).

is about two feet across and 0.06 foot deep. Most of the curved stones are somewhat flexible.

The headstones are all oriented with their intermediate dimension trending about north; the curve as commonly faces east as west and as commonly away from as toward the grave, evidently not controlled by tilting toward a settling grave. Curvature or its amount are not apparently related to the age of the headstones. One of the youngest stones examined, dated 1884, is curved 5 degrees. Minor peaks in the total number of stones and in the number of curved stones occur in those dated about 1830 and 1850; possibly this marble was more readily available at those times.

POSSIBLE EXPLANATIONS

The bending of the headstones seems to be controlled by the physical characteristics of the rock, and not by external factors. Of the three possible explanations given below, none is completely satisfactory, but the writers feel that the first one given, one involving partial disaggregation of mineral grains, is most reasonable.

(1) Bending may be due to the effect of gravity on the upper part of a stone originally not truly vertical, as essentially interlocking calcite grains are partly disaggregated and loosened by differential solution of interstitial material. Once disaggregation is begun the retention of moisture is favored,

and continued solution at pressure points in the loose, interlocking network of calcite grains would result in continued bending and eventual failure by collapse when the network had been so weakened that it could no longer support the upper part of the stone. This mechanism explains the flexibility of the stones but does not satisfactorily explain the saucer-like warping.

(2) Deformation of the stones may be due to an unloading or stress release process.

(3) Curving and warping of the headstones may involve differential expansion associated with weathering of some minerals. This explanation is weakened by the nearly monomineralic makeup of most of the stones.

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REVIEW

Ordovician Cephalopod Fauna of Baffin Island; by A. K. MILLER, WALTER YOUNGQUIST, and CHARLES COLLINSON, containing a study of the *Ordovician Trilobites from Silliman's Fossil Mount;* by H. B. WHITTINGTON. Geological Society of America Memoir 62. P. 234; 20 figs., 63 pls. New York, 1954 ($4.75).—A volume issued in the impressive format of a Geological Society Memoir is born with a prestige quite independent of its intrinsic merits and therefore requires and deserves more careful review than the same material might demand in more humble form. Let me say at once that this volume is an extremely valuable addition to our knowledge of the Arctic Ordovician faunas, and sincere gratitude is due the authors for their thorough study and presentation of their material. Having said that, and most heartily, there are one or two points which I think require comment.

It should be kept in mind that this volume records the results of three days' collecting from a talus slope. The authors assure us that this procedure (which made possible larger collections than could have otherwise been gathered in the time available) was justified because the 350 feet of beds contained a single marine fauna. Confidence in this assurance is somewhat shaken by a later statement that most of the large cephalopods and *Maclurites* came from the upper 50 feet. It would be unfair to tax the authors for not doing in three days what would require weeks, but the possibility must be held open that collecting by beds might still reveal a faunal succession within the beds exposed in the Mount.

For the detailed descriptions of the cephalopods, trilobites, and graptolites which make up the bulk of the volume there can be only admiration, and no detailed comments are necessary. That the work has been done by Miller's group, Whittington, and Decker is sufficient guarantee of its reliability.

Most interest will center on the discussion of the age of the fauna, a discussion which summarizes all published information on cognate faunas in the Arctic and elsewhere. The authors conclude that the beds are Upper Ordovician, on what seems to me very little evidence. Let us examine each group of Silliman fossils in turn.

Corals.—Few corals were collected, and they are noncommittal as to age. *Halysites* occurs commonly as low as the Black River, and *Calapoecia* is typically a Black River genus. (Since *Calapoecia* is shown in the Ordovician correlation chart as a Richmond fossil, this dictum must be supported. Miller et al. quote Cox' statement (1936) that *Calapoecia* is "highly characteristic of post-Trenton and Richmond strata," but this statement needs examination. Cox recognized only one species of *Calapoecia*, *C. canadensis*, the type locality of which is near Ottawa in beds of Black River age. Of that there is no question. Confusion enters because Cox, unable to locate the type specimens, selected a neotype from slightly younger beds at Paquette Rapids, some 75 miles up the Ottawa River from the type locality. Then, since this specimen was silicified, Cox described the internal structure from specimens from the Richmond of Southern Ontario and Kentucky (1936, p. 10). Under the circumstances it is not surprising that he found little difference between "typical" *C. canadensis* and Richmond forms of the genus. This confusion was doubly confounded since, under a misapprehension, he included in his "Richmond"

material specimens from the Black River at Lake Saint John, Quebec (p. 35). In short, although Cox' work is admirable biologically, it is so muddled stratigraphically as to be of little use. Bassler has previously suggested (1950, p. 276) that Cox' work needs clarification.)

Graptolites.—Only three species could be identified by Decker, but one of them is important, for it is a *Dicranograptus* and suggests that the fauna containing it should be earlier than Ashgillian. (Ruedemann has described species of *Dicranograptus* from the "Upper Ordovician" of the western United States, but in each case these occurrences are in beds of disputed age and cannot be used to prove a post-Caradocian occurrence of the genus.)

Echinoderms.—The few species collected from Frobisher Bay tell us little about possible correlation of the beds. Most of the genera represented are found in the Trenton and also in the Maquoketa. The identification of *Lichenocrinus affinis* by Schuchert cannot be given much weight. Not only are these structures of doubtful taxonomic value but Schuchert's identification was made before the Fentons described several species from the Black River and Trenton.

Bryozoans.—Again there is little to go on. Miller et al. quote from Bassler's work of 1911 that he could identify positvely two species but omit his further statement that these species were characteristic of the Nematopora bed [Prosser] in Minnesota. That Bassler now thinks the Arctic beds Richmond is interesting, but citing this opinion in connection with discussion of the bryozoans leaves an impression that the species listed are Richmond species, which is not so.

Brachiopods.—A considerable brachiopod fauna is listed, with Cloud's opinion that the indicated age is Trenton, apart from the presence of *Austinella.* (I have inverted the statement in Miller et al., but I think not unfairly.) The significance of the species of *Austinella* is not easy to assess. This genus has been reported only rarely, and then always from Upper Ordovician beds, but I am not sure that it should be considered absent from older beds until their brachiopod faunas have been reworked in modern terms. Thus, *Dinorthis iphigenia media* (Wilson, 1946, p. 41) from the Trenton has a brachial valve quite as convex as the Arctic *Austinella*, and the convexity of this valve seems to be the chief external distinction of the genus. Certainly the other brachiopods are strikingly Trenton in aspect, especially if the *Rhynchotremas* be compared not with the small Sherman Fall *R. increbescens* but with the Cobourg *R. intermedia* (Wilson, 1932, p. 400).

Ostracods.—Warthin lists the species found in the fauna and gives his opinion that the age should be Trenton.

Trilobites.—Whittington, on the basis of his thorough study of this group, gives the definite opinion that they suggest correlation with the Trenton, perhaps even Lower Trenton. Like Warthin, he concedes that the age might possibly be post-Trenton, but only if some hitherto unknown fauna is involved. In both cases it is clear that on the basis of the groups they studied Warthin and Whittington prefer a Trenton age for the material.

Cephalopods.—Here Miller is on home ground, and here I think the implied correlation of the beds shows most clearly. The authors accept Kay's

demonstration that the Red River faunas of Manitoba and the Stewartville of Iowa are so similar as to suggest correlation, but they seem unable to accept Kay's further correlation of the Stewartville with the Cobourg in the East. In support of their stand they cite opinions that the Stewartville may be younger than had been thought, and Upper Ordovician. This reassignment of the Stewartville is difficult to refute in the Mississippi Valley, where the Maquoketa rests on the Stewartville-Dubuque sequence, and the latter could conceivably be immediately pre-Maquoketa in age.

Here I think Miller et al. underestimate the importance of the Terrebonne fauna. For if the position of the Stewartville is equivocal, that of the Terrebonne is clear. The Terrebonne (Clark, 1952, p. 74 f.) is overlain by the Lachine formation of the Utica group, that by the Lorraine, and that in turn by the Richmond. The Terrebonne is indisputably the upper part of the Trenton limestone sequence at Montreal, it underlies beds of Utica age, and it contains the Red River-Stewartville cephalopods. Flower has described species of *Apsidoceras, Lambeoceras, Fremontoceras, Winnipegoceras,* large *Probillingsites* and *Diestoceras* from it, and has announced the presence of *Westonoceras* and *Charactoceras* (1952, p. 25 f.). Flower's work needs no defense by me, but, as Miller et al. seem to question the validity of this fauna, it might be well to emphasize that Flower knows it not only from laboratory study but also in the field, having visited with me almost all the Terrebonne localities, and having collected as well large faunas from the Lower and Middle Trenton of Quebec, which provide a context for evaluation of the Terrebonne fauna itself. The Black River beds at Lake Saint John (Miller et al., p. 33) have nothing to do with the Terrebonne fauna.

In summary, we have in this Silliman Mount fauna as described much to suggest that the beds (if they are all of one age) may be Trenton and nothing to demand that they must be younger but the brachiopod genus *Austinella.* Undoubtedly, as more work is done in the Arctic, the picture will become clearer. Let us not blur our vision of it by the prejudgment that an Upper Ordovician age is established for this fauna so well collected and described by Miller and his associates.

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